Origin and geomorphology of Dover Strait and southern North Sea palaeovalleys and palaeo-depressions.

Oorsprong en geomorfologie van de Straat van Dover Straat en de paleovalleien en paleodepressies van de zuidelijke Noordzee.

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Research highlights

- **The opening of the Dover Strait was not due to tectonic forcing or structurally controlled. Our study demonstrates that the activity of the various tectonic structures traversing the Dover Strait has been very low during the Quaternary Period. Earthquakes of magnitudes 6–7, although possible, have been rare and deformation along faults negligible.**

- **The morphological analysis of the Fosses Dangeard indicates that they were incised by waterfalls, possibly during the Elsterian glacial maximum. This demonstrates the presence of a lake in the southern North Sea at least once during the Pleistocene Epoch. Our study suggests that the ridge damming that lake at the Dover Strait was subjected to waterfall headward erosion, which led to the, possibly sudden, breach of the ridge, thus initiating the opening of the Dover Strait. This may have resulted in the generation of a megaflood in the English Channel, further eroding the Fosses Dangeard and incising part of the Channel palaeovalleys. The morphology and infills of the Fosses Dangeard indicate the occurrence of only one phase of plunge-pool erosion. The formation of a second lake dammed at the Dover Strait is thus unlikely. This confirms previous hypotheses postulating that ice-marginal lakes formed in the southern North Sea during subsequent Pleistocene lowstands were more localized. Internal erosional surfaces identified in the infills of the Fosses Dangeard suggest the occurrence of at least two other episodes of intense fluvial or flood erosion in the span between the formation of the Fosses Dangeard and the last phase of intense valley incision that imprinted the seafloor.**

- **The Lobourg Channel was mainly shaped by 3 episodes of intense valley incision, attesting to the occurrence of high-magnitude flood flows or high-energy fluvial erosions in the Dover Strait. The morphology of the inner channel carved during the last phase of valley incision and its associated erosional features strongly resemble those found in megaflood-eroded terrains, indicating that at least that one was produced by a megaflood. The geophysical data show that this megaflood occurred once the Fosses Dangeard were formed and infilled.**

- **The Axial and Lobourg Channels are different parts of the same palaeovalley, into which the Rhine–Meuse and Thames palaeo-drainage systems seem to have converged in Late Saalian and Weichselian times. This palaeovalley is joined by a major N–S-oriented palaeovalley (i.e. the North Axial Channel). The location and orientation of that channel is consistent with a major palaeo-drainage system, which, according to palaeogeographic reconstructions, flowed southward from the Danish–German continental shelf during the Last Glacial Maximum. Correlations among inner channels carved within the North Axial, Axial and Lobourg channels suggest that the megaflood event responsible for the last phase of valley incision in the Dover Strait was most likely produced 30–18 ka ago.**

- **Our study also corroborates the southward diversion of the eastern end of the Thames–Medway palaeo-river, which possibly occurred during the Elsterian glacial maximum. The geomorphology of the Outer Thames Estuary indicates that the ice sheets may have extended over that area during the Elsterian glaciation, covering part of the Thames–Medway watercourse and inducing its southward deviation.**

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Figure 7.15. Possible palaeovalleys and deeps formed during the Elsterian glaciation identified in this study. Ice-sheet margin at Elsterian glacial maximum proposed by previous studies and hypothetic ice expansion along the OTE (this study) are indicated..

Figure 7.16. Possible palaeovalleys (blue areas) and scarps (dashed red line) carved during the Saalian glaciation interpreted from the data discussed in this study. Ice-sheet southern margin at Saalian glacial maximum (black line labelled IM – MIS 6) is indicated. RMNE–MIS 6: Saalian Rhine–Meuse northern edge; RMSE–MIS 6: Saalian Rhine–Meuse southern edge (see Figure 7.1).

Figure 7.17. Possible palaeovalleys (blue areas) incised during the Weichselian glacial maximum interpreted from the data discussed in this study. Ice-sheet margin at Weichselian glacial maximum (black line labelled IM – MIS 2) is indicated (see Figure 7.1). Labels are explained in previous figures and main text.

Figure 7.18. Palaeovalleys (blue areas) representing the last phase of valley incision identified in the study area according to interpretations of the data discussed in this study. Ice-sheet southern margins

at Weichselian glacial maximum (black line labelled IM – MIS 2) and at ~15 ka BP (yellow line labelled IM> 15 ka; Dove et al., 2017) are indicated. Labels are explained in previous figures and main text.

Figure 8.1. Sketch showing the hypothetical hydrographic configuration before (a) and after (b) the breach of the Weald–Artois ridge. Lake extent in (a) after Gibbard and Cohen (2015). Lake extent and palaeo-river watercourses in (b) hypothesized by the author of this study. The various palaeovalleys of the Channel palaeovalley system are indicated; these are: LC: Lobourg Channel; SB: South Basserelle palaeovalley; NPV: Northern palaeovalleys; MPV: Median palaeovalley. Mw: Medway palaeo-river. TI: hypothetical tongue of ice extending southward across the Outer Thames Estuary (OTE).

Figure 8.2. Sketch showing an alternative hypothetical scenario proposed for the southward diversion of the Elsterian Thames–Medway palaeo-river system across the OTE; i.e. before and/or after the ice-sheets merged across the central–northern North Sea.

Figure 8.3. a) Sketch showing the hydrographic network during the Drenthe Stage of the Saalian glacial maximum before the draining of the southern North Sea lake. Modified from Gibbard (1988), Gibbard (1995) and Gibbard and Cohen (2015). b) Hypothetical hydrographic network after the breach of the land-bridge damming the ice-marginal lake based on the geomorphology of the English Channel and southern North Sea seafloors. River watercourses are hypothesized by the author of this dissertation from combinations of the results of this study with reconstructions from Gibbard (1988), Gibbard (1995), Cohen et al. (2014) and Mellett et al. (2013).

Figure 8.4. a) Sketch showing the hydrographic network during the Warthe Stage of the Saalian glacial maximum, supposing a continuous river system without development of lakes in the southern North Sea. River watercourses are hypothesized by the author of this dissertation from combinations of the results of this study with reconstructions from Gibbard (1988), Gibbard (1995), Cohen et al. (2014) and Mellett et al. (2013). See previous figures for other labels.

Figure 8.5. Sketch depicting the hypothetical hydrographic configuration during the two possibly Last valley incisions along the Axial–Lobourg Channel. a) Before the Lic–Aic–NAic incision; b) after the Lic–Aic–NAic incision.

Table 1.1. Land-based chronostratigraphic terminology used for major Quaternary glacial and interglacial periods in northwestern Europe and their correlation with Marine Isotope Stages, absolute timing and Geological timescale (modified from Gibbard and Cohen, 2008

Table 3.1. Seismic reflection datasets available for the present study. Lille U.: Lille 1 University; RCMG: Renard Center of Marine Geology; ROB: Royal Observatory of Belgium; Emu Ltd–IECS-UH: Emu Ltd and the Institute of Estuarine Coastal Studies (University of Hull).

Table 3.2. Bathymetric datasets available for the present study. BCS: Belgian Continental Shelf; DCS: Dutch Continental Shelf; GSB: Gabbard Sandbank Bathymetry; OTE REC: Outer Thames Estuary Regional Environmental Characterization; EC REC: East Coast Regional Environmental Characterization; MBES: Multibeam echosoundings; SBES: Single-beam echosoundings; SSS: side-scan sonar; HRSRP: High-resolution seismic-reflection profiles; Lille U.: Lille 1 University; RCMG: Renard Center of Marine Geology; ROB: Royal Observatory of Belgium; FHS: Flemish Hydrographic Service; MCA: Maritime and Coastguard Agency; Multiple Inst.: Multiple institutions; GGOW: Greater Gabbard Offshore Wind; IC: Imperial College London. See Table 3.1 for other labels

Table 6.1. Bedrock units outcropping in the Dover Strait.

Table 6.2. Geophysical datasets available for the present study. Lille U.: Lille University; RCMG: Renard Center of Marine Geology (Ghent University); ROB: Royal Observatory of Belgium; MCA: British Maritime & Coastguard Agency; UKHO: United Kingdom Hydrographic Office; SHOM: Service Hydrographique et Océanographique de la Marine; IC: Imperial College London; SC: Single-channel seismic-reflection data; MC: Multi-channel seismic-reflection data; SBES: single-beam bathymetric data; MBES: Multibeam bathymetric data.

Table 6.3. Possible correlations between erosional events and geomorphologic features identified in the Dover Strait. The possible timing of their occurrence is also indicated.

Abbreviations

AC: Axial Channel

AD: Anno Domini; i.e. year after birth of Christ

Aic: Axial Inner Channel

BE: Belgium

BCS: Belgian Continental Shelf

BDCSB: Belgian–Dutch Continental Shelf Banks

BP: Before present

CDP: Common Depth Point

Ch: channel

CMP: Common Mid Points

CT: Channel Tunnel

DCS: Dutch Continental Shelf

DE: Germany

DK: Denmark

DS: Dover Strait

DTM: Digital Terrain Model

EC: English Channel

EMODnet: European Marine Observation and Data Network

FBB: Flemish Bight Banks

FD: Fosses Dangeard

Fz: fault zone

FR: France

GB: Great Britain

GEBCO: General Bathymetric Chart of the Oceans

IE: Ireland

IGD: Inner Gabbard Deeps

IM: ice maximum or maximum extent of an ice sheet.

Ka: kiloannum; i.e. ten thousand years

L: lineations

LC: Lobourg Channel

Lic: Lobourg Inner Channel

LAT: Lowest Astronomical Tide

Ma: Mega Annum; i.e. 1 Million years.

MBES: multibeam bathymetric echosoundings.

MPV: Median palaeovalley

MIS: Marine Isotope

MSK intensity

NAC: North Axial Channel

NB: Norfolk Banks

NL: The Netherlands

NMO: Normal move-out

NO: Norway

NPV: Northern palaeovalleys

NS: North Sea

OTE: Outer Thames Estuary

Pv: palaeovalley

Pvs: palaeovalley system

RCMG: Renard Center of Marine Geology (Department of sciences, Faculty of science, Ghent

University).

REC: Regional Environmental Characterization

RM or R–M: Rhine-Meuse palaeo-river system

ROB: Royal Observatory of Belgium

SB: South Basserelle palaeovalley

Sc: scour

Sb: sandbank (in this study, Sb is synonym of TSR)

SBES: single-beam echo soundings

Sd, also as D: sand dune

SHOM: Hydrographic and Oceanographic Service of the French Marine

SEG-Y: Society of Exploration Geophysicists 'Y' format

SRTM: Shuttle Radar Topography Mission (i.e. Topographic digital elevation models)

SU: Seismic Unix

TE: Thames Estuary

TEB: Thames Estuary Banks

TSR: Tidal sand ridge

TWT: Two-Way Time, equivalent to Two-Way-Travel Time (TWTT)

UTM: Universal Transverse Mercator coordinate system

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Samenvatting

Het Engelse Kanaal en de zuidelijke Noordzee herbergen spectaculaire paleovalleien en -depressies in de zeebodem, die het bewijs vormen voor intense terrestrische erosie die ooit in het verleden plaatsvond. Paleogeografische en paleoklimatologische reconstructies tonen aan dat deze gebieden inderdaad boven de zeespiegel lagen tijdens de Midden tot Laat Pleistocene ijstijden. Niettegenstaande zijn de geogenetische processen die deze paleovalleien en-depressies gevormd hebben nog grotendeels onbekend.

De Fosses Dangaerd zijn een van belangrijkste en meest controversiële erosiestructuren in de overgangszone van de zuidelijke Noordzee naar het Engelse Kanaal. Samen met honderden kilometer lange paleovalleien in het noorden (e.g. Axiale Kanaal) en het zuidwesten (i.e. de Paleovalleien in het Kanaal) zijn zij van cruciaal belang voor het ontwerpen en uitwerken van paleogeografische reconstructies voor de hele regio. Vandaag de dag wordt de vorming van deze grote landschapselementen toegewezen aan een van de belangrijkste paleogeografische veranderingen die plaatsvonden tijdens het Pleistoceen, i.e. het doorbreken en permanent openen van de Straat van Dover.

Paleogeografische reconstructies tonen aan dat de Straat van Dover ooit een aaneengesloten topografisch barrière was tussen het Verenigd Koninkrijk en Frankrijk. Deze topografische barrière, of de Weald-Artois-anticline, vormde de afscheiding tussen de zuidelijke Noordzee en het Engelse Kanaal tot ca. 450 ka BP tijdens het Elsteriaan Glaciaal Maximum. Tijdens die glaciale periode waren de Brits-Ierse en Fennoscandische ijskappen tot een geheel samengesmolten boven de centrale Noordzee, waardoor een deel van de zuidelijke Noordzee geïsoleerd werd van de noordelijke Atlantische Oceaan. Ten gevolge van de ijskapvorming vond ook uitgebreid glacio-isostasie plaats in het hele gebied van de zuidelijke Noordzee. Verschillende auteurs stellen voor dat deze configuratie tot gevolg had dat het deel ten zuiden van de ijskap onder water kwam te staan ten gevolge van de vorming van een proglaciaal meer. Door de glacio-isostatische effecten van de ijskap vond grote subsidentie plaats in Nederland waardoor de oevers van dit meer tot diep in Nederland en België reikten. Dit meer werd in het noorden door de aaneengesloten ijskap afgedamd terwijl in het zuiden de Weald-Artoisanticline fungeerde als een barrière. Het bestaan van dit meer is vandaag de dag grotendeels aanvaard; echter, tot op heden kon hiervoor nog geen sluitend bewijs worden aangebracht.

Op basis van de hierboven vernoemde paleogeografische landschapselementen hebben sommige auteurs een hypothese opgesteld die stelt dat de Fosses Dangaerd het resultaat zijn van watervallen toen het marginaal ijsmeer de Weald-Artoisanticline begon te overstijgen ten tijde van het Elsteriaan Glaciaal Maximum. Of deze initiële opening van de Straat van Dover het resultaat is van progressieve fluviatiele erosie van de Weald-Artoisanticline of van een plotse catastrofale doorbraak is op dit moment nog een discussiepunt. Aan de ene kant hebben verschillende auteurs voorgesteld dat, als de Weald-Artoisanticline plots doorbreekt, een zeer snel einde zou gemaakt worden aan de waterval-fase en dus een megavloed voortgebracht zou hebben in het Engelse Kanaal. Deze megavloed zou dan de Fosses Dangaerd verder uitgeschuurd hebben en een deel van het Kanaal Paleovallei systeem gevormd hebben in het Engelse Kanaal.

Aan de andere kant zijn er auteurs die voorstellen dat de Straat van Dover (en dus ook de Fosses Dangaerd) uitgeschuurd werd door progressieve fluviatiele erosie. Deze studies associëren de paleodepressies met grote begraven en blootgestelde paleovalleien die doorheen de Straat van Dover lopen (i.e. het Lobourg Kanaal en de Paleovallei in het Kanaal).

Een laatste hypothese stelt dat de opening van de Straat van Dover, de vorming van de Fosses Dangaerd en het Lobourg Kanaal samen het resultaat zijn van een Neogeen-Quartaire reactivering van Variscische tektonische structuren die de Straat van Dover kruisen t.g.v. een combinatie van episodische isostatische bewegingen gecombineerd met tektoniek. Deze stelling wordt ondersteund door de locatie van de Fosses Dangaerd die inderdaad uitgeschuurd zijn binnen een grote Variscische breukzone (i.e. de Noord Artois Schuifzone). De breukzone werd in het verleden geassocieerd met enkele historische aardbevingen van gematigde magnitude alsook enkele lokale geologische *offsets* van Quartaire sedimentaire eenheden die duiden op een Quartaire activiteit. Desondanks is deze mogelijks Quartaire activiteit nooit gekarakteriseerd. Bovendien is de continuatie van deze structuur op zee grotendeels onbekend.

Sinds het einde van het Elsteriaan Glaciaal zijn de zuidelijke Noordzee en het Engelse Kanaal nog twee keer boven de zeespiegel komen te staan, tijdens het Midden tot Laat Pleistoceen, i.e. de Saaliaan (350-130 ka BP) en Weichseliaan Glacialen (110-12 ka BP), en nog drie maal onder water komen te staan tijdens de interglacialen die hier telkens op volgden. Deze klimatologische en paleogeografische veranderingen hadden tot resultaat dat grote veranderingen van erosieve/sedimentologische aard plaatsvonden binnen dit gebied. Deze hebben mogelijk geleid tot significante veranderingen in de morfologie en sedimentatie van verschillende erosiestructuren gevormd tijdens het voorgaande Elsteriaan Glaciaal, alsook de vorming van enkele nieuwe. Meer specifiek waren er enkele lokale en paleogeografische configuraties tijdens de Saaliaan en Weichseliaan Glaciale Maxima, respectievelijk 175-140 ka en 30-25 ka geleden, die blijkbaar een significante bijdrage geleverd hebben tot de vorming van de hedendaagse geomorfologie van het Engelse Kanaal en de zuidelijke Noordzee. Gedurende deze periodes zouden de Brits-Ierse en Fennoscandische ijskappen opnieuw samengesmolten zijn boven de centrale Noordzee, waardoor de verbinding zuidelijke Noordzee-noordelijke Atlantische Oceaan wederom afgesloten werd. Paleogeografische reconstructies suggereren dat in deze configuratie opnieuw een proglaciaal meer tot stand kwam in de zuidelijke Noordzee tijdens het Saaliaan Glaciaal Maximum. Sedimentologische en biostratigrafische data tonen aan dat dit meer zich tientallen kilometer tot in Nederland uitstrekte, doch het reikte niet tot aan de westelijke Belgische kustvlakte. De ligging van dit meer in het noordelijke deel van het ijsvrije zuidelijke Noordzee is geassocieerd met de aanwezigheid van een landbrug dat zich uitstrekte tussen East Anglia en noordwest België. Deze landbrug wordt tevens verantwoordelijk geacht voor het afdammen van dit meer in het zuidwesten. De lagere hoogteligging van het meer zou dan het gevolg zijn van glacio-isostasie langsheen de zuidelijke rand van de ijskap. De algemene kenmerken van deze landbrug zijn op dit moment echter niet goed gekend. Paleogeografische reconstructies suggereren dat het verwijderd werd, ten dele, voor 160 ka BP, wat leidde tot de vorming van een groot drainage-systeem in de zuidelijke Noordzee, dat ook de Rijn, Maas en de Thames omvatte. Deze paleorivieren lijken immers te convergeren in een diep uitgesneden paleovallei in de zuidelijke Noordzee, het Axiale Kanaal, dat lijkt verbonden te zijn met de paleovalleien in het Kanaal stroomafwaarts van de Straat van Dover d.m.v. het Lobourg Kanaal.

Een gelijkaardig paleorivier-systeem lijkt tot stand te zijn gekomen in de zuidelijke Noordzee-Engelse Kanaal tijden het Weichseliaan Glaciaal Maximum. Hoewel in dit geval het systeem aangevuld werd door een zeer groot riviersysteem dat gevormd werd door de confluentie van de ijsmarginale Duitse Rivieren of door de uitstroom van een ijsmarginaal meer dat gevoed werd door deze Duitse rivieren.

Deze paleogeografische reconstructies zijn algemeen aanvaard. Doch, geen specifieke geomorfologische onderzoeken zijn tot vandaag de dag uitgevoerd in de zuidelijke Noordzee en de Straat van Dover die de verschillende Midden Pleistocene hydrografische/geografische configuraties, zoals hierboven voorgesteld, uittesten. Bijgevolg blijven vele vragen onbeantwoord. Zo weten we bijvoorbeeld nog niet wanneer en hoe de Fosses Dangaerd en de Paleovalleien in het Kanaal tot stand zijn gekomen. Op dit moment is zeer weinig geweten over de verschillende tektonische structuren die deze doorkruisen en welke rol ze gespeeld hebben in hun vorming of de opening van de Straat van Dover. Bovendien is de configuratie van de drainage-systemen in de zuidelijke Noordzee gedurende de verschillende Midden-Laat Pleistocene mariene laagstanden zeer slecht gekend. De meeste van deze modellen blijven tot op de dag van vandaag zeer hypothetisch.

Om enkele van deze vragen te kunnen oplossen en de validiteit van de hierboven toegelichte hypothesen te evalueren hebben we een geomorfologische studie uitgevoerd van de verschillende paleovalleien en paleodepressies die waargenomen zijn binnen dit gebied. We concentreren ons in deze studie op het Lobourg Kanaal, de Fosses Dangaerd in de Straat van Dover, en het Axiale Kanaal. Bij deze laatste wordt ook de relatie met de Thames, Rijn en Maas, alsook de ijsmarginale Duitse paleorivieren naderbij bestudeerd. Om dit te doen hebben we voor het studiegebied een grote hoeveelheid bathymetrische en reflectie-seismische data verzameld en gecompileerd.

De geomorfologische en geologische analyses van de geofysische data van de Straat van Dover demonstreren dat de opening ervan of de vorming van de Fosses Dangaerd inderdaad een tektonische component bezat. Echter hun vorming of morfologie zijn niet significant beïnvloed door deze kruisende tektonische structuren. De 3D morfologische analyse van de Fosses Dangaerd tonen aan dat ze hoogst waarschijnlijk gevormd zijn door noordwaarts migrerende watervallen die het bestaan van een meer in de zuidelijke Noordzee aantonen dat afgedamd werd nabij de Straat van Dover ergens tijdens het Pleistoceen. De morfologie en de opvulling van de Fosses Dangaerd suggereren ook het voorkomen van minstens één episode van waterval erosie dat abrupt ten einde kwam. Dit is consistent met de catastrofale doorbraak van de Weald-Artoisanticline, hetgeen op zijn beurt een megavloed in het Engelse Kanaal heeft veroorzaakt.

De bovenste sedimentlagen van de Fosses Dangaerd zijn weggeschuurd tijdens de laatste fase van vallei-insnijding in het Lobourg Kanaal. Het Lobourg Kanaal werd gevormd door drie fasen van vallei-insnijding. De ouderdom van de overige twee fasen en hun relatie met de verschillende episodes van erosie en afzetting in de Fosses Dangaerd zijn minder goed gekend, maar ze dateren waarschijnlijk van na de initiele uitschuring van de Fosses Dangaerd. Zowel de opvulling als de zeebodem-geomorfologie van de Fosses Dangaerd suggereren het voorkomen van intense fluvio-hydraulische en/of megavloed erosie in de Straat van Dover gevolgd door een insnijding en de gedeeltelijke opvulling van de Fosses Dangaerd. Zo hebben twee van de vele geïdentificeerde erosieoppervlakken in de centraal gelegen Fosses morfologische kenmerken die vergelijkbaar zijn met deze van depressies uitgeschuurd door intense fluvio-hydraulische erosie van rivieren en/of megavloeden. Bovendien zijn de drie vallei-insnijdingen die het Lobourg Kanaal gevormd hebben, en in het bijzonder de recentste, gelijkaardig aan erosiestructuren waargenomen in door megavloed geërodeerde landschappen. Deze studie stelt daarom voor dat verschillende dergelijke externe fluviatiele en/of megavloed erosiegebeurtenissen in de Straat van Dover hebben plaatsgevonden sinds de vorming van de Fosses Dangaerd. Het tijdstip van deze gebeurtenissen is echter niet goed gekend. Met de huidige data zijn we enkel in staat om voorzichtig de timing van de initiële opening van de Straat van Dover en de recentste paleovallei-insnijding van het Lobourg Kanaal af te bakenen. Sedimentaire gegevens die verzameld werden in de Golf van Biscaje tonen aan dat twee zeer hoog-energetische pulsen van fluvio-hydraulische energie langsheen de Paleorivieren van het Kanaal hebben plaatsgevonden ten tijde van de Elsteriaan en Saaliaan Glaciale Maxima. Op basis hiervan stellen we voor dat de oudste van deze fluviatiele pulsen overeenstemt met het doorbreken van de Weald-Artois-anticline en de initiële opening van de Straat van Dover. Aan de andere kant tonen de geomorfologie van de zuidelijke Noordzee en de Straat van Dover aan dat het Lobourg Kanaal de zuidelijke continuatie is van het Axiale Kanaal waarin alle grote paleorivier-systemen waargenomen in de zuidelijke Noordzee convergeerden. Deze omvatten niet alleen de paleovalleien van de Rijn-Maas en Thames maar ook een grote paleovallei verder noordwaarts (i.e. het Noordelijke Axiale Kanaal). Op basis van vergelijkingen met recente paleogeografische reconstructies stellen we hier voor dat het Noordelijke Axiale Kanaal gevormd werd tijden het Weichseliaan Glaciaal Maximum, door ofwel de uitstroom van een proglaciaal meer gesitueerd in het gedeelde Deens-Duitse continentaal plat of door de convergentie van ijsmarginale Duitse paleorivieren die afweken naar het zuiden. Correlaties van de intern uitgesneden geulen in de Noordelijke Axiale, Axiale en Lobourg Kanalen suggereren dat een megavloed-erosie, geïdentificeerd langsheen het Lobourg Kanaal, mogelijks plaatsvond tussen 30-18 ka.

Ook al beschikken we over geen data omtrent het voorkomen van andere erosie-gebeurtenissen die de Fosses Dangaerd en het Lobourg Kanaal gevormd hebben, onze studie toont wel aan dat ze plaatsvonden ergens tussen de vorming van de Fosses Dangaerd en de laatste vallei-insnijding langsheen het Lobourg Kanaal. Bijgevolg, als de hypotheses zoals hierboven toegelicht correct zijn, zijn deze insnijdingen gevormd na de opvulling van de Fosses Dangaerd en het Lobourg Kanaal tussen de Gaciale Maxima van het Elsteriaan en Weicheliaan. Hierdoor is het eerder waarschijnlijk dat sommige of alle van deze insnijdingen al gevormd werden tijden het Saaliaan Glaciaal. Dit ondersteunt het voorkomen van significante fluviatiele vloeden langsheen de Paleorivieren van het Kanaal tijdens deze glaciatie, mogelijks, gegenereerd door de drainage van het Saaliaan ijsmarginaal meer dat grote delen van de zuidelijke Noordzee onder water zette.

Summary

The English Channel and southern North Sea exhibit spectacular palaeovalleys and palaeo-depressions carved into their seafloors, attesting to several phases of intense subaerial erosion. Palaeogeographic and palaeo-climatic reconstructions indicate that these areas were indeed emerged during major Middle and Late Pleistocene glaciations. However, the processes that led to the formation of these apparent palaeo-fluvial systems are still largely unknown.

Probably the most important for palaeogeographic reconstructions, and controversial, erosional features located in the southern North Sea – English Channel area are the kilometer-scale buried depressions known as the Fosses Dangeard, and a series of deeply incised, hundreds-of-kilometers long palaeovalleys carved into the seafloors of the English Channel (i.e. the Channel palaeovalleys) and in the southern North Sea (e.g. the Axial Channel). Indeed, the formation of these features has been associated with one of the most important geographic modifications that took place during the Pleistocene Epoch; i.e. the opening of the Dover Strait.

Palaeogeographic reconstructions indicate that the Dover Strait was closed by a ridge known as the Weald-Artois ridge, which appears to have separated the southern North Sea from the English Channel, at least until the Elsterian glacial maximum (~450 ka BP). At that time, the British–Irish and Fennoscandian ice-sheets merged across the central North Sea, isolating part of the southern North Sea from the North Atlantic Ocean and inducing subsidence in the North Sea basin due to glacial isostatic depression. Several authors have proposed that this configuration led to the inundation of the unglaciated part of the southern North Sea by a major proglacial lake. This lake may have extended over large parts of the Netherlands and Belgium owing to the possibly larger subsidence induced along the southern ice-sheet margin than further south. It was purportedly dammed in the north by the merged Fennoscandian–Irish–British ice sheets, in the east and south by continental Europe, in the west by southern Great Britain and in the southwest by the Weald–Artois ridge. The presence of that proglacial lake at Elsterian glacial maximum is widely accepted; however, no conclusive evidence of its existence has been found yet and its extent is currently debated.

Based on the particular palaeogeographic configuration during the Elsterian glacial maximum described above, some authors have hypothesized that the Fosses Dangeard are the result of plunge-pool erosion at the base of waterfalls, which formed as the southern North Sea ice-marginal lake overtopped the Weald-Artois ridge. According to that hypothesis, waterfall headward erosion would have eventually led to the opening of the Dover Strait. However, there is no consensus on whether that happened due to progressive fluvial erosion or due to a sudden, catastrophic, breach of the ridge. On the one hand, several authors have proposed that the Weald–Artois ridge breached suddenly, generating a megaflood in the eastern English Channel. This megaflood would have further incised the Fosses Dangeard and carved part of the Channel palaeovalley system in the English Channel. On the other hand, some authors have proposed that the breach in the Weald–Artois ridge was opened by progressive fluvial erosion of the cataract wall. These studies associate the formation of the Fosses Dangeard with the incision of a prominent palaeovalley imprinted on the Dover Strait (i.e. the Lobourg Channel of the Channel palaeovalley system).

Alternatively to the hypotheses outlined above, it has also been proposed that the opening of the Dover Strait, the excavation of the Fosses Dangeard and the incision of the Lobourg Channel might have been induced and/or controlled by a Neogene–Quaternary reactivation of Variscan tectonic structures traversing the Dover Strait. According to that hypothesis, these structures may have been reactivated during the Quaternary by a combination of tectonic forcing and episodic isostatic depression and rebound. The Fosses Dangeard are indeed carved within a major Variscan fault zone (i.e. the North Artois Shear Zone). This fault zone has been associated with some moderate-magnitude historical earthquakes, as well as with local deformation of Quaternary sedimentary units. However, the possible Quaternary activity of these faults and the potential impact this may have had on the opening of the Dover Strait have never been characterized. In addition, the structure of the offshore continuation of the North Artois Shear Zone remains largely unknown.

Since the end of the Elsterian Glaciation, the southern North Sea and English Channel have been emerged again twice in the Middle and Late Pleistocene Epoch, i.e. during the Saalian (350–130 ka BP) and Weichselian Glaciations (110–12 Ka BP), and submerged three times, i.e. during the interglacial periods that followed each Middle–Late Pleistocene glaciation. These climatic and palaeogeographic changes resulted in several major modifications of the erosional/depositional settings of this area, possibly inducing significant changes in the morphology and infills of the various erosional features formed during the Elsterian Glaciation, as well as the formation of new ones. In particular, local climatic and palaeogeographic settings during the Saalian and Weichselian glacial maxima appear to have significantly contributed to shape the present-day geomorphology of the English Channel and southern North Sea regions. During those periods, the Irish–British and Fennoscandian ice sheets merged across the central North Sea at 175–140 ka BP and 30–19 ka BP, thus isolating the southern North Sea from the North Atlantic Ocean.

Palaeogeographic reconstructions suggest that another ice-marginal lake may have formed in the southern North Sea during the Saalian glacial maximum. Sedimentological and biostratigraphic data indicate that this lake extended tens of kilometers into the Netherlands. However, sedimentological data collected from the Yser River indicate that it did not extend over the Belgian western coastal plain. This is consistent with recent palaeogeographic reconstructions, which proposed that this lake was restricted to the northern part of the unglaciated area of the southern North Sea. The restriction of that lake to the north is currently explained by the presence of a land-bridge extending between East Anglia and northeastern Belgium, and by the possible relatively lower elevation of the ice-sheet southern margin induced by glacial isostatic depression. The nature of the land-bridge that purportedly dammed the lake in the south is currently unknown. Palaeogeographic reconstructions suggest that it was removed, at least partially, before 150 ka BP, which resulted in the establishment of a major palaeo-drainage system in the southern North Sea. This palaeo-drainage system comprised the Rhine, Meuse and Thames palaeo-rivers. These palaeo-rivers appear to have converged into a prominent channel, commonly referred to as the "Axial Channel", which appears to connect with the Channel palaeovalley system across the Dover Strait through the Lobourg Channel. A similar palaeo-drainage system seems to have been established in the southern North Sea – English Channel during the Weichselian glacial maximum. However, in that case, this system may have been joined by a significant paleo-river formed by the confluence of southwestward diverted German Rivers or by the main outflow of an ice-marginal lake.

Most of these palaeogeographic reconstructions are widely accepted. However, no specific geomorphological investigation has been conducted so far in the southern North Sea and

Dover Strait to test the validity of the different Middle–Late Pleistocene hydrographic/geographic configurations proposed for these areas. Hence, many questions remain unanswered. For instance, we still do not know how and when the Fosses Dangeard and Lobourg Channel formed. Little is also known about the role that the various tectonic structures traversing those features may have played on their formation or on the opening of the Strait. In addition, the configuration of the southern North Sea hydrographic network during the various Middle–Late Pleistocene marine lowstands is poorly constrained. Most of the models proposed for the evolution of the major rivers running across the southern North Sea remain indeed hypothetical. In order to solve some of these questions and evaluate the validity of the various hypotheses proposed for the Pleistocene evolution of the Dover Strait and southern North Sea, we have undertaken a geomorphological characterization of the various palaeovalleys and palaeo-depressions carved into these areas. Particularly, we focus on the Lobourg Channel and Fosses Dangeard, in the Dover Strait, and on the Axial Channel and its relationship with the Thames, Rhine, Meuse, and southward diverted German palaeo-rivers, in the southern North Sea. For that purpose, we have gathered a large dataset of bathymetric and seismic reflection data, covering large parts of those areas.

The geomorphological and geological analyses of the geophysical data collected from the Dover Strait demonstrate that neither the opening of the Dover Strait nor the scouring of the Fosses Dangeard were caused by tectonic forcing. Neither were their formation or morphology significantly controlled by the tectonic structures traversing that area. The three-dimensional morphological analysis of the Fosses Dangeard indicates that these features were most likely formed by northward retreating waterfalls. This implies the presence of a lake in the southern North Sea dammed at the Dover Strait at least once during the Pleistocene Epoch. The morphology and infill of the Fosses Dangeard also suggest the occurrence of only one episode of waterfall erosion, which appears to have ended suddenly. This is consistent with a catastrophic breach of the Weald–Artois ridge, which may have generated a megaflood in the English Channel. However, the timing of these events is poorly constrained. Sedimentary data collected from the Gulf of Biscay show the occurrence of massive fluvial discharges from the Channel palaeo-river, which appear to have happened during the Elsterian and Saalian glacial maxima. Based on this, we propose that the oldest of these fluvial discharges corresponds to the breach of the Weald-Artois ridge and initial opening of the

Dover Strait.

The present-day Lobourg Channel was formed by three major valley incisions, the last one of which occurred once the Fosses Dangeard were formed and fully infilled. The relative timing of the previous two valley incisions is less well constrained, although they were also formed following the formation of the Fosses Dangeard.

Both the infills of the Fosses Dangeard and the geomorphology of the seafloor suggest the occurrence of intense fluvial and/or high-magnitude flood erosion in the Dover Strait following the incision and partial infilling of the Fosses Dangeard. For instance, two of the many internal erosional surfaces identified in the infills of the central Fosses present similar morphological characteristics to scours carved by rapid, highly-erosive rivers and/or by high-magnitude flooding. In addition, the three valley incisions that shaped the Lobourg Channel, especially the youngest one, generated erosional features similar to those observed in megaflood-eroded terrains.

The geomorphology of the southern North Sea and Dover Strait confirms that the Lobourg Channel is the southward continuation of the Axial Channel, into which all major palaeovalley systems identified in the southern North Sea appear to converge. These include not only the Rhine–Meuse and Thames palaeovalleys, but also a major palaeovalley located further north (i.e. the North Axial Channel). Based on comparisons with recent palaeogeographic reconstructions, we propose that the North Axial Channel was incised either by the outflow of a proglacial lake, which may have been formed in the Danish–German continental shelf 30–19 ka ago, or by converging German palaeo-rivers that may have been diverted toward the Dover Strait between 30–18 ka BP. Correlations between inner channels carved within the North Axial Channel, the Axial Channel and the Lobourg Channel suggest that the megaflood responsible for the youngest valley incision identified in the Lobourg Channel occurred within the 30–18 ka interval or during later stages of the Weichselian glaciation.

The ages of the internal erosional surfaces and other valley incisions indicating high-energy fluvial and/or high-magnitude flood erosion identified, respectively, in the infills of the Fosses Dangeard and in the Lobourg Channel are unknown. Nevertheless, our investigation shows that they happened in the span between the formation of the Fosses Dangeard and the youngest major valley incision identified in the Lobourg Channel. Hence, if the hypotheses proposed above are correct, these erosional features were carved between the Elsterian and Weichselian glacial maxima. It is thus likely that some or all of them were incised during the Saalian glaciation. This is consistent with the occurrence of significant fluvial discharges along the Channel palaeo-river during that glaciation identified in previous studies.

The geomorphological study of the southern North Sea has also provided important information on the evolution of the Thames palaeo-river during the various Middle–Late Pleistocene glaciations. In particular, the spatial distribution of buried and exposed palaeovalleys in the Thames Estuary and Outer Thames Estuary corroborate a southward migration of the Thames–Medway palaeo-river as proposed in previous studies. In addition, we have found geomorphological evidence supporting the possible southward diversion of the eastern part of this palaeo-river system during the Elsterian glaciation. This southward diversion may be related to the expansion of the Elsterian ice-sheets over that area, as suggested by the morphology and spatial distribution of the various palaeovalleys and palaeo-depressions incised in the Outer Thames Estuary.

To sum up, despite the lack of absolute dating, our investigation permits to validate/reject a series of hypotheses on which palaeogeographic reconstructions of the southern North Sea and eastern English Channel are based. This study thus provides important new information on the hydrographic and palaeogeographic evolution of northwestern Europe during major Middle–Late Pleistocene glaciations, as well as on the processes that led to the present-day geography and geomorphology of the southern North Sea and English Channel.

Chapter 1

Introduction

Chapter 1 – Introduction

From the first bathymetric charts constrained in the late 1800’s to the present-day geophysical investigations, the geomorphology and Quaternary geology of the English Channel and southern North Sea areas have picked the interest of the scientific community. The submarine landscapes of these areas are indeed spectacular. They comprise: hundreds-of-kilometers long, tens-of-kilometers wide, bedrock-incised palaeovalleys; tens-of-meters deep, kilometer-scale buried palaeo-depressions; and tens-of-kilometers long elongated sandbanks (Figure 1.1; e.g. Caston, 1972; Destombes et al., 1975; Henriet et al., 1989; Dyer and Huntley, 1999; James et al., 2002; Antoine et al., 2003; Gupta et al., 2007; Sejrup et al., 2009; Hijma et al.,2012; Mellett et al., 2013; Collier et al., 2015). Over the years and with the improvement of the resolution of offshore geophysical techniques, some of these geomorphological features have been attributed to (and used as proof of) the different major climatic changes that have affected northwestern Europe during the Quaternary Period. However, the formation and age of many of these features remain largely unknown, as little reliable dating is available and high-resolution geophysical/geological data are still lacking from large parts of these areas.

Two of the most significant and controversial erosional features located in the southern North Sea – English Channel area are the Fosses Dangeard and the Channel palaeovalley system, also known as the Channel/Manche River (Figure 1.1 and Figure 1.2; Destombes et al., 1975; Smith, 1985; Mellett et al., 2013; Collier et al., 2015). The formation of these features appears to be linked to one of the most significant palaeogeographic modifications that took place during the Pleistocene Epoch; i.e. the opening of the Dover Strait (e.g. Smith, 1981; Gibbard, 1995; Gupta et al., 2007; Toucanne et al., 2009; Gibbard and Cohen, 2015). However, the origin of these features and their relationship with the opening of the Dover Strait are uncertain. Their geomorphology and interrelationship are also poorly constrained. Especially, the three-dimensional geometry of the Fosses Dangeard has never been imaged.

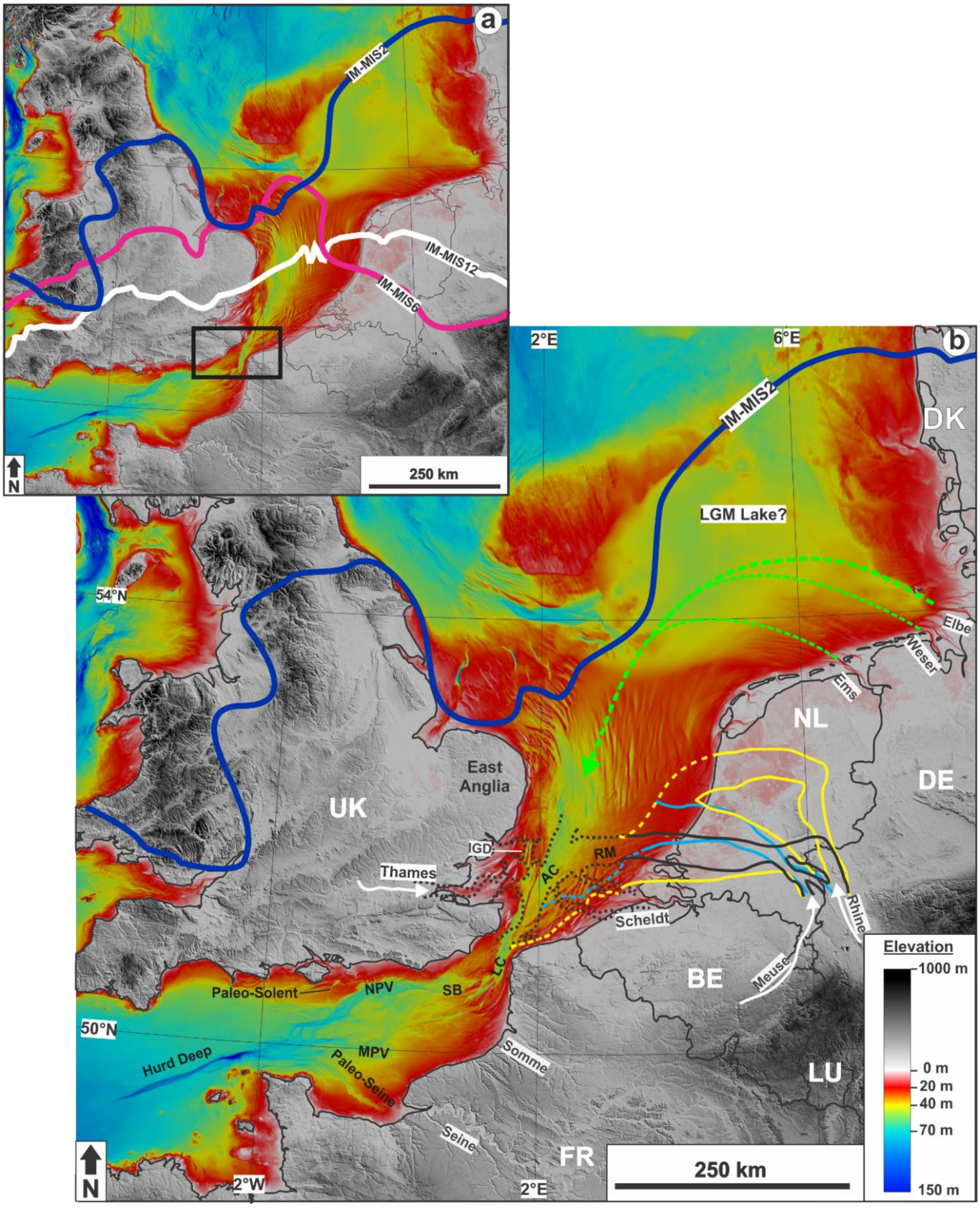

Figure 1.1. a) North Sea southern maximum extents of ice sheets during Elsterian, Saalian and Weichselian glacial maxima (pink, white and dark blue lines, respectively), after Ehlers and Gibbard (2004), Lee et al. (2012) and Sejrup et al. (2016). Black rectangle: area shown in Figure 1.2. b) Main exposed and buried palaeovalleys imprinted on the seafloors of the English Channel and southern North Sea (after Hijma et al., 2012; Mellett et al., 2013). Dotted and dashed lines: hypotheses; continues lines: based on geological and/or geomorphological evidence. Light blue lines, Yellow and black lines represent the palaeo-watercourses of the Rhine–Meuse palaeo-river at 150 ka BP, at 80–35 ka BP and at 30–25 ka BP, respectively (Busschers et al., 2007; Peeters et al., 2015; Hijma et al., 2012). IM: ice maximum extent; RM: possible continuation of the Rhine–Meuse palaeovalley

at 30-25 ka (Hijma et al., 2012); AC: Axial Channel; IGD: Inner Gabbard Deeps. Major palaeovalleys composing the Channel palaeovalley system are labelled as: LC: Lobourg Channel; SB: South Basserelle palaeovalley; NPV: Northern palaeovalleys; MPV: Median palaeovalley; and Hurd Deep. Light green dashed lines indicate the hypothetical southwestward deviation of northwestern German rivers (after Toucanne et al., 2010). "LGM Lake" indicates the possible location of an ice-marginal lake that may have occupied that area 30–19 ka ago (Sejrup et al., 2016). Bathymetry: EMODnet dataset (230 m cell size). Topographic information: SRTM worldwide elevation data (3-arc-second resolution). Projections: UTM, zone 31N. Horizontal datum: WGS84; SRTM vertical datum: EGM96; Bathymetric vertical datum: LAT.

Another palaeogeographic issue that is still unsolved is that of the actual paths taken by the various north European palaeo-drainage systems traversing the southern North Sea during Pleistocene marine lowstands. Several hypotheses exist on the different paths the major palaeo-river systems may have taken at those periods (e.g. Hijma et al., 2012). However, few of these are based on actual offshore geomorphological evidence and, those that are (e.g. Henriet et al., 1989; Liu et al., 1992), are either too localized or focused on specific erosional features. Regional-scale geomorphological investigations of buried and exposed palaeovalleys and palaeo-depressions have never been undertaken prior to this study. Rather, existing regional hypotheses on the spatial distribution of palaeo-rivers in the southern North Sea and their interrelationship during the different Middle–Late Pleistocene glaciations are based on combinations of onshore palaeo-drainage reconstructions (e.g. Lewis et al., 2004; Busschers et al., 2007) and/or local offshore geomorphological/geological studies (e.g. Liu et al., 1992; Mathys, 2009) with reconstructions of the ice-sheet extents during those periods (e.g. Ehlers and Gibbard, 2004; Carr et al., 2006; Sejrup et al., 2009, Lee et al., 2012).

The present study is intended to characterize the morphology of, and interrelationship between, the various palaeovalleys and palaeo-depressions carved into the seafloors of the Dover Strait and southern North Sea. The main objective of this study is to better understand how and when these features formed. The formation of these features is indeed key to unravel how the Dover Strait opened, as well as to better understand how it evolved since its formation to its present-day configuration. In addition, a more detailed geomorphological characterization of the submarine southern North Sea and Dover Strait areas is imperative to better constrain the characteristics of the drainage systems that traversed these areas during the various Middle–Late Pleistocene glacial periods. Such study may also provide the geomorphological evidence lacking in previous palaeogeographic reconstructions to validate the hypotheses on which they are based.

In the first part of this study, we characterize the morphology and infill of major palaeo-depressions and palaeovalleys carved in the Dover Strait. In particular, we focus on the enigmatic tens-of-meters deep Fosses Dangeard, and on the prominent palaeovalley known as the Lobourg Channel, which purportedly links the main palaeovalleys imprinted on the southern North Sea with the Channel palaeovalleys (Figure 1.1). In this first part of the study, we combine seismic reflection and bathymetric data to analyze the activity of the various tectonic structures traversing the Dover Strait and their possible relationship with the formation of the various palaeo-depressions and palaeovalleys incised in that area. This analysis is complemented with detailed characterizations of the morphologies and infills of the Fosses Dangeard, as well as with a detailed geomorphological analysis of the Lobourg Channel and its associated erosional features. In the second part of this dissertation, we investigate the continuation of the Lobourg Channel into the southern North Sea and its relationship with other major palaeovalley systems carved in that area (see Figure 1.1). We also describe in detail the palaeovalleys carved along the Thames Estuary and Outer Thames Estuary areas, and those possibly incised by southwestward diverted major German river systems during the Last Glacial Maximum. Finally, we compare our geomorphological analyses with regional and local palaeogeographic reconstructions in an attempt to unravel their minimum age, as well as to assess the validity of previous palaeogeographic and hydrographic models in the light of our findings.

1.1 Chronostratigraphy

Before introducing the different geomorphological features that we will describe and characterize in this dissertation, we will briefly introduce the meaning and equivalences between the various terminologies typically used in northwestern Europe to refer to the different Quaternary stages, sub-stages and climatic cycles.

In marine geology, it is widely accepted to use the Marine Isotope Stage (MIS) timescale to define the interval of time during which a geological/climatic event took place. This timescale is deduced from oxygen isotope data from deep sea sediment samples, which reflects changes in the global temperature over time. Stages with even numbers have high levels of ^{18}O and represent cold glacial periods; whereas odd-numbered stages have low levels of ^{18}O and represent warm interglacial intervals. The MIS timescale is therefore commonly used to define the duration of the various Quaternary glacial and interglacial stages and sub-stages.

The MIS timescale defines cold and warm intervals at global scale. However, the characteristics of the various Quaternary glacial and interglacial periods are usually based on local, land-based studies. This has derived in the use of completely different nomenclatures to refer to those cycles in different parts of the world. For instance, MIS 6 corresponds to the glaciation that took place between 200–130 ka BP (i.e. thousands of years before present). That glaciation is known as "Saalian" in Northern Europe, "Wisconsin" in North America and "Wolstonian" in Great Britain. Another issue that may lead to confusion is that the glacial maximum – defined as the maximum extent of the ice-sheets during a glacial period – of each Pleistocene glaciation did not occur at the same time in different parts of the world.

In order to avoid misunderstandings, we show in Table 1.1 the various land-based chronostratigraphic terminologies used to define major Quaternary climatic cycles in northwestern Europe and their equivalences with Marine Isotope Stages, absolute timing and the Quaternary time scale (see also Gibbard and Cohen, 2008). We also indicate the time intervals during which the Fennoscandian and Irish–British ice sheets merged across the North Sea. These correspond to the various Middle–Late Pleistocene glacial maxima. However, during the Weichselian glaciation, the ice sheets may have remained merged across the North Sea during a longer time interval than the one commonly associated with its last glacial maximum (i.e. 30–22 ka BP; e.g. Graham et al., 2007; Graham et al., 2011). Indeed, Sejrup et al. (2016) have proposed that the Fennoscandian and Irish–British ice sheets did not separate from each other until 19–18.5 ka BP. Based on that, we assume 19 ka BP as the minimum age for the time interval during which the ice sheets merged across the North Sea during the Last Glacial Maximum.

Table 1.1 only includes glacial and interglacial intervals for the timeframe during which the erosional/depositional features discussed in this study were purportedly formed; i.e. from 400 ka to the present. Note that, in this dissertation, we use the various chronologic terms introduced in Table 1.1 as synonym.

Glacial/interglacial periods		Inter/ Glacial	MIS	Period (ka BP)	ice sheets merged across NS (ka BP – MIS)	Epoch	Stage
Northern Europe	Great Britain						
	Flandrian	interglacial	1	present – 12		Holocene	
Weichselian/ Last Glacial period (Age)	Devensian	glacial	2 – 5d	12–110	30–19* (MIS 2); ~70 (MIS 4)?	Pleistocene	Late
Eemian	Ipswichian	interglacial	5e	110–130			
Saalian	Wolstonian/ Gipping	glacial	6 – 10	130–374	175–155; 150–140 (both MIS 6)		Middle
Holsteinian	Hoxnian	interglacial	11	374–424			
Elsterian	Anglian	glacial	12	424–478	~450		

Table 1.1. Land-based chronostratigraphic terminology used for major Quaternary glacial and interglacial periods in northwestern Europe and their correlation with Marine Isotope Stages, absolute timing and Geological timescale (modified from Gibbard and Cohen, 2008; Sejrup et al., 2016*). Time intervals during which Fennoscandian and Irish–British ice sheets merged across the North Sea are indicated.

1.2 The Fosses Dangeard and the Lobourg Channel

1.2.1 The Fosses Dangeard

The Fosses Dangeard are a set of sediment-filled paleo-depressions incised tens-of-meters into bedrock and distributed over a 20 km long belt, extending along the width of the Dover Strait (Figure 1.2; Destombes et al., 1975; Smith, 1985; James et al., 2002). These buried palaeo-depressions were first discovered during the geophysical site surveys undertaken for the construction of the Channel Tunnel, and described for the first time by Destombes et al. (1975). However, descriptions of their morphology and infills in that and subsequent studies have remained scarce (Smith, 1985; James et al., 2002). For instance, these studies provided no data supporting the isopach maps they shown. A more recent study (e.g. Oggioni, 2013) provided geophysical evidence of the morphology and infill of the central palaeo-depressions of the Fosses Dangeard. However, that investigation focused only on the central buried deeps and did not interpret the entire complex of those palaeo-depressions. Consequently, despite the many studies undertaken in this area, the 3-dimensional morphology of these palaeo-depressions remains largely unknown.

The age and nature of the sediments infilling the Fosses Dangeard are also unknown.

Currently, only one borehole exists and it did not penetrate the entire infill (Destombes et al., 1975). In addition, it sampled sediments infilling only one of the multiple palaeo-depressions composing this feature, which prevents establishing whether the sampled sediments represent a local facies or whether all depressions contain a similar infill.

Initially, the formation of the Fosses Dangeard were attributed to glacial erosion (Destombes et al., 1975), and tentatively associated with the southward expansion of ice sheets during the Saalian glacial maximum (MIS 6; 175–140 ka BP). Nevertheless, that hypothesis has been discarded, as sedimentary and geomorphological evidence indicates that none of the Middle Pleistocene ice sheets reached the Dover Strait (e.g. Clark et al., 2004; Ehlers and Gibbard, 2004; Carr et al., 2006; Sejrup et al., 2009, Lee et al., 2012).

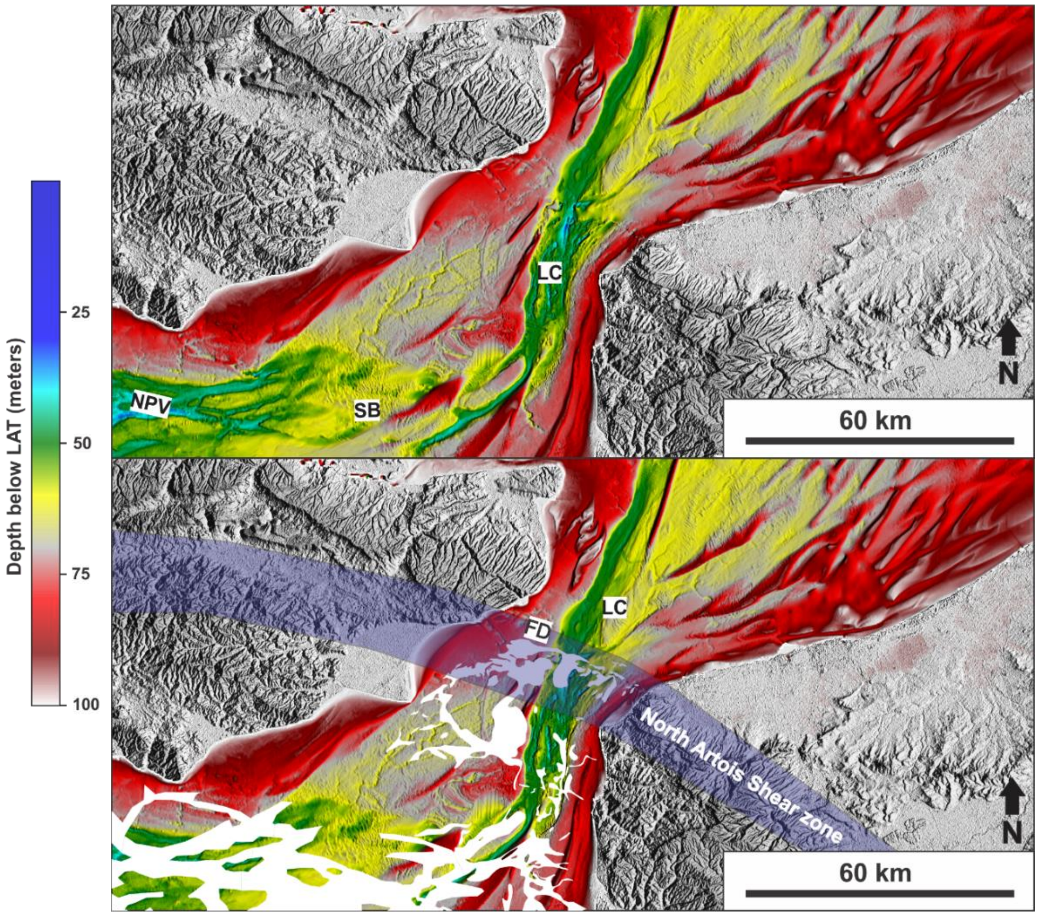

Figure 1.2. Exposed and buried palaeovalleys and palaeo-depressions carved in the Dover Strait. Above, bathymetric map showing the Lobourg Channel incision across the Dover Strait. Below, buried palaeovalleys and palaeo-depressions (white areas) superimposed on the bathymetry (after maps from James et al., 2002). Position and plan-view morphology of the Fosses Dangeard (FD) are indicated (after Destombes et al., 1975; Smith, 1985; James et al., 2002). Violet belt: approximate location of the North Artois Shear Zone. Bathymetry: EMODnet dataset (230 m cell size).

Alternatively, Smith (1985) interpreted these paleo-depressions as fossil plunge pools carved at the base of waterfalls. Smith (1985) theorized that the ridge over which these waterfalls flowed was eventually breached, initiating the opening of the Dover Strait. For this to happen, the southern North Sea should have contained a lake dammed at the Dover Strait. This is consistent with palaeogeographic reconstructions, which postulate that the Dover Strait was closed by a ridge known as the Weald-Artois ridge. This ridge purportedly connected northern France with southeastern England until the Elsterian glacial maximum, i.e. ~450 ka ago (e.g. Gibbard, 1995; Toucanne et al., 2009a; Gibbard and Cohen, 2015; Gibbard and Lewin, 2016). This, added to the blockage of the northern routes of palaeo-rivers by the coalescence of the Irish–British and Fennoscandian ice sheets across the central North Sea, induced the inundation of the unglaciated part of the southern North Sea. The breach of the Dover Strait dam and the emptying of the southern North Sea proglacial lake during the Elsterian glaciation is also consistent with sedimentary information collected from the Gulf of Biscay. Sediments sampled in that area suggest the occurrence of massive fluvial discharges from the Channel palaeo-river at the Elsterian glacial maximum, thus suggesting that the Dover Strait may have been opened, at least partially, at some point during that stage (Toucanne et al., 2009a). However, there is no consensus on how this happened. On the one hand, there are those who propose that the opening of the Dover Strait occurred progressively. According to these authors, the Fosses Dangeard are the result of progressive fluvial erosion within a major river system; i.e. the Channel palaeo-river (e.g. Hamblin et al., 1992; Westaway and Bridgland, 2010; Mellett et al., 2013). On the other hand, there are those who, based on several extreme-erosional features discovered in the English Channel (Smith, 1985; Gupta et al., 2007; Collier et al., 2015; Oggioni, 2013), propose that the Dover Strait dam breached suddenly, generating a high-magnitude flood flow (also known as a megaflood) in the English Channel. According to these authors, this event incised further the Fosses Dangeard and carved some of the Channel Palaeovalleys in the English Channel. This hypothesis is also supported by coarse gravels containing occasional northern erratics and Fennoscandian heavy minerals found at the base of channels in northern France (Roep et al., 1975; Gibbard, 1995).

Finally, it has also been proposed that these palaeo-depressions and the palaeovalleys incised in the Dover Strait were largely controlled by tectonic forcing and isostatic adjustments (e.g. Colbeaux et al., 1980; Van Vliet-Lanoë et al., 2004). The Dover Strait is indeed traversed by a regional fault-and-fold system known as the North Artois Shear Zone (Figure 1.2). This tectonic structure appears to have been reactivated several times since its formation, during the Variscan orogeny (370–259 Ma ago), up to the present (Vandycke et al., 1988; Bergerat & Vandycke, 1994; Van Vliet-Lanoë et al., 2002a,b; Van Vliet-Lanoë et al., 2004; Mansy et al., 2003; Minguely et al., 2010). According to Van Vliet-Lanoë et al. (1998; 2002a,b; 2004), the Dover Strait went through a series of closing-and-opening episodes during the Quaternary Period due to a reactivation of the North Artois Shear Zone by a combination of glacio-isostatic adjustments and tectonic stress. These authors proposed that it was not until 160–130 ka ago (Saalian–Eemian transition) that the Dover Strait opened definitively. Evidence of Quaternary faulting along the North Artois Shear Zone is however scarce. The only traces of Quaternary activity along this fault system are minor local offsets identified on some of its segments (e.g. Colbeaux et al., 1980) and the location of a historical (i.e. the 1580 Dover Strait earthquake) and a recent (i.e. the 2007 Folkestone earthquake) earthquake of moderate magnitude along its offshore prolongation (e.g. Camelbeeck et al., 2007; Ottemöller et al. 2009). The potentially important role that the Weald-Artois Shear Zone may have had (e.g. structural/tectonic control of the erosion) on the opening of the Strait and the incision of the Fosses Dangeard has however never been investigated in detail.

The formation of the Fosses Dangeard will be addressed in Chapters 4 and 5 of this dissertation, in which we investigate in detail the morphology and infills of these palaeo-depressions, and the activity of the tectonic features traversing them, in order to evaluate the validity of the different hypotheses postulated for their formation.

1.2.2 The Lobourg Channel

The Lobourg Channel is a broad NE–SW-oriented palaeovalley extending across the Dover Strait and cut predominantly into bedrock (Figure 1.2; James et al., 2002; Mellett et al., 2013; Oggioni, 2013; Collier et al., 2015). This palaeovalley forms part of the Channel palaeovalley system, a complex anastomosing system of infilled and unfilled valleys that can be mapped continuously from the Dover Strait to the Hurd Deep (Figure 1.1; e.g. Antoine et al., 2003; Lericolais et al., 2003; Mellett et al., 2013; Collier et al., 2015). Both its morphology and its sedimentary infill indicate that this palaeovalley system was formed by fluvial erosion (Mellett et al., 2012; Mellett et al., 2013; Collier et al., 2015). However, there is no consensus on the exact processes that steered the erosion. Moreover, the timing and evolution of the various valley incisions that shaped these palaeovalleys are still poorly constrained. Several authors have associated the incision of this palaeovalley system with progressive erosion along a major palaeo-river, into which all northwestern European and southern and southeastern British palaeo-rivers discharged during Middle–Late Pleistocene marine lowstands (Antoine et al., 2003; Lericolais et al., 2003; Mellett et al., 2012; Mellett et al., 2013). On the other hand, recent studies have identified several erosional features carved within these palaeovalleys similar to those found in megaflood-eroded terrains, indicating that the Channel palaeovalley system, or some of its channels, may have been carved by one or several episodes of high-magnitude catastrophic flooding (Gupta et al., 2007; Oggioni, 2013; Collier et al., 2015).

Despite the many studies conducted on the geomorphology of the Channel palaeovalleys, it is still unknown whether the Fosses Dangeard formed contemporarily to the incision of the Lobourg Channel. In addition, the evolution of the Lobourg Channel, from its first incision to its present-day geomorphology, and its relationship with other palaeovalleys incised in the southern North Sea and Dover Strait have never been characterized. This palaeovalley should indeed be the result of several phases of fluvial valley incision and flooding, as the Dover Strait was emerged over thousands of years during the last 3 major glaciations. In addition, this area has been subjected to 3 major marine transgressions. Tidal erosion and submarine currents might thus have contributed to shape the morphology of this palaeovalley too.

In Chapters 5 and 6, we conduct a detailed analysis of the geomorphology of the submarine Dover Strait in order to shed light on the origin and evolution of the Lobourg Channel, as well as its relationship with the incision of the Fosses Dangeard and opening of the Strait. In Chapter 7, we explore the continuation of this channel across the southern North Sea and its relationship with other major palaeovalleys incised in that area (Figure 1.1).

1.3 The southern North Sea palaeovalleys

Present-day paleo-geographic reconstructions postulate that the Fennoscandian and British–Irish ice sheets merged across the North Sea during the last three major Pleistocene glacial maxima (Figure 1.1; e.g. Ehlers and Gibbard, 2004; Sejrup et al., 2009, Lee et al., 2012; Sejrup et al., 2016). This caused the isolation of the then emerged southern North Sea from the North Atlantic Ocean, inducing major changes in the drainage systems running through that area (see Figure 2.5 in Chapter 2; e.g. Gibbard, 1995; Gibbard and Cohen, 2015). This might have led to the formation of a proglacial lake in the southern North Sea during the Elsterian and part of the Saalian glacial maxima, and the development of a major palaeo-river system during Late Saalian and Late Weichselian glaciations (e.g. Gibbard, 1995; Hijma et al., 2012; Gibbard and Cohen, 2015). This river system appears to have comprised all rivers running into the southern North Sea (e.g. the Rhine, Meuse and Thames Rivers), which were diverted toward the Dover Strait due to the merged ice sheet blocking their previous northern routes. That palaeo-river system thus connected with the Channel palaeo-river, forming a continuous river system that traversed the southern North Sea and English Channel and discharged into the Atlantic Ocean.

Even though the evolution of the southern North Sea hydrographic network during the various Pleistocene marine lowstands has been the subject of several studies (e.g. Figure 1.1), many questions remain unsolved. Firstly, the hydrographic configuration of palaeo-drainage systems traversing the southern North Sea during Saalian and Weichselian glacial maxima is poorly constrained. Secondly, there is no consensus on the drainage pattern of a series of German rivers running into the unglaciated part of the southern North Sea during the Last Glacial Maximum (Figure 1.1). On the one hand, some authors suggest that these rivers formed an ice-marginal lake in the German–Danish continental shelf of the southern North Sea (Sejrup et al., 2016; Patton et al., 2017). These authors hypothesized that such a lake would have outflowed at its southwestern end, forming a stream that joined with the main Rhine–Meuse–Thames palaeo-river system further to the south; i.e. in the Axial Channel (Sejrup et al., 2016). On the other hand, other authors have proposed that these rivers converged into a large river system, which also ran southwestward and converged with the Rhine–Meuse–Thames palaeo-river system in the Axial Channel (Figure 1.1; Toucanne et al., 2010; Toucanne et al., 2015).

Finally, a series of palaeovalleys and palaeo-depressions are carved in the Outer Thames Estuary. The morphology of these palaeo-depressions has never been studied in detail. In addition, very little offshore geomorphological evidence has been presented to support the hypothesis postulated by, among others, Bridgland and D'Olier (1995) on the southward diversion of the eastern part of the Thames–Medway palaeo-river and later southward migration of its watercourse during the Elsterian Glaciation.

The hydrographic configuration of the southern North Sea during Middle–Late Pleistocene glacial maxima is investigated in Chapter 7. There, we perform detailed geomorphological analyses of the Outer Thames Estuary and the part of the southern North Sea that was free of ice during the three last Pleistocene glaciations in order to characterize the morphology and evolution of the various palaeovalleys and palaeo-depressions incised in that area.

1.4 Organization of this dissertation

This dissertation consists of 9 Chapters:

- Chapter 1, the current Chapter, provides a general introduction to the study area and research questions addressed in this dissertation.
- In Chapter 2, the general setting of the study area is presented.
- Chapter 3 describes the methods used for this study as well as the available datasets.
- In Chapter 4, we investigate the activity of the faults traversing the Dover Strait and their relationship with the formation of the Fosses Dangeard. This chapter is intended to assess whether a Quaternary tectonic – or tectonically controlled – opening of the Dover Strait is plausible, as well as the possible structural control the folding and faulting may have had on the formation of the Fosses Dangeard. This Chapter has been published, in a slightly modified version as: "*García-Moreno, D., Verbeeck, K., Camelbeeck, T., De Batist, M., Oggioni, F., Zurita Hurtado, O., Versteeg, W., Jomard, H., Collier, S.J., Gupta, S., Trentesaux, A., and Vanneste, K., 2015. Fault activity in the Epicentral area of the 1580 dover-strait/Pas-de-Calais earthquake (North-Western Europe). Geophysical Journal International, v. 201, p. 528–542*." Note that the contributions of the author of this dissertation to this and subsequent multi-authored published and unpublished chapters are detailed below the abstracts of each chapter.
- In Chapter 5, we analyze and interpret the morphology of the Fosses Dangeard and the most significant erosional features associated with the incision of the Lobourg Channel in order to test the validity of the hypotheses proposed for their formation. This Chapter has been published, in a slightly modified version, as "*Gupta, S., Collier, J.S., Garcia-Moreno, D., Oggioni, F., Trentesaux, A., Vanneste, K., De Batist, M., Camelbeeck, T., Potter, G., Van Vliet-Lanoë, B., Arthur, J.C., 2017, Two-stage opening of the Dover Strait and the origin of island Britain. Nature Communications, 8, doi:*

10.1038/ncomms15101."

- Chapter 6 is an extension of Chapter 5, where we describe and analyze in much detail the morphologies and infills of the Fosses Dangeard and Lobourg Channel. With this extended and more detailed geomorphological analysis, we aim to further demonstrate the hypotheses postulated in Chapter 5 on the origin of these features. This Chapter also addresses the evolution of these features since their formation; i.e. the possible number, nature and timing of the various scouring-and-infilling episodes that shaped their present-day morphology. This Chapter will be submitted for publication in November–December 2017, as "*Garcia-Moreno D., Gupta, S., Collier, J.S., Oggioni, F., Vanneste, K., Trentesaux, A., Verbeeck, K., Versteeg, W., Jomard, H., Camelbeeck, T., De Batist, M., 2017–2018, Characterization of late Quaternary palaeo-landscapes in the Dover Strait and their implication for the opening of the Strait. Quaternary Sciences Review*."
- In Chapter 7, we characterize the morphology and interrelationship of the various Pleistocene palaeovalleys imprinted on the southern North Sea. With this, we aim to test the different palaeogeographic models proposed for the evolution of the major southern North Sea drainage systems during the Pleistocene glaciations. In this Chapter, we also make correlations between the different phases of erosion identified in the Dover Strait and those imprinted on the southern North Sea in order to better estimate their origin, palaeogeographic context and relative age.
- In Chapter 8, we combine all investigations in a general discussion, in which we revise the hypotheses postulated in previous palaeogeographic reconstruction based on our findings.
- And, finally, in Chapter 9 the overall conclusions of this study are presented.

References

Antoine, P., Coutard, J-P., Gibbard, P., Hallegouet, B., Lautridou, J-P., and Ozouf, J-C., 2003, The Pleistocene rivers of the English Channel region, Journal of Quaternary Science, v.18, p. 227–243.

Bridgland, D.R., and D'Olier, B., 1995, The Pleistocene evolution of the Thames and Rhine drainage systems in the southern North Sea Basin, in: Preece, R. C. (ed.), 1995, Island Britain: a Quaternary perspective Geological Society Special Publication, v. 96, p. 27-45

Busschers, F. S., Kasse, C., Van Balen, R. T., Vandenberghe, J., Cohen, K. M., Weerts, H. J. T., Wallinga, J., Johns, C., Cleveringa, P., and Bunnik, F. P. M., 2007, Late Pleistocene evolution of the Rhine-Meuse system in the southern North Sea basin: imprints of climate change, sea-level oscillation and glacio-isostacy, Quaternary Science Reviews, v. 26, p. 3216–3248.

Camelbeeck, T., Vanneste, K., Alexandre, P., Verbeeck, K., Petermans, T., Rosset, P., Everaerts, M., Warnant, R., and Van Camp, M., 2007, Relevance of active faulting and seismicity studies to assessments of long-term earthquake activity and maximum magnitude in intraplate northwest Europe, between the Lower Rhine Embayment and the North Sea, The Geological Society of America, v. 425, p. 193–224.

Carr, S.J., Holmes, R., van der Meer, J.J.M., Rose, J., 2006, The last glacial maximum in the North Sea basin; micromorphological evidence of extensive glaciation, Journal of Quaternary Science, v. 21, p. 131–153.

Caston, V.N.D., 1972, Linear sand banks in the southern North Sea. Sedimentology, v. 18, p. 63–78.

Collier, J. S., Oggioni, F., Gupta, S., García-Moreno, D., Trentesaux, A., De Batist, M., 2015, Streamlined islands and the English Channel megaflood hypothesis, Global and Planetary Change, v. 135, p. 190–206.

Colbeaux, J-P., Depuis, C., Robasznki, F., Auffret, J-P., Haesaerts, P., and Somme, J., 1980, Le détroit du Pas-de-Calais : Un Element dans la tectonique de blocs de l'Europe Nord-occidentale. Bulletin d'information des géologues du bassin de Paris, v. 17, p. 41–54.

Destombes, J. P., Shephardthorn, E. R., Redding, J. H., and Morzadec-kerfourn, M. T., 1975, Buried valley system in the Strait of Dover, Philosophical Transactions of the Royal Society A-Mathematical Physical and Engineering Sciences, v. 279, p. 243–256.

Dyer, K.R., and Huntley, D.A., 1999, The origin, classification and modelling of sand banks and ridges, Continental Shelf Research, v.19, p. 1285–1330.

Ehlers, J., and Gibbard, P. L. (Eds.), 2004, Quaternary glaciations-extent and chronology: part I: Europe. Developments in Quaternary Science, vol. 2, Elsevier, Amsterdam.

Graham, A.G.C., Lonergan, L., Stoker, M.S., 2007, Evidence for Late Pleistocene ice stream activity in the Witch Ground basin, central North Sea, from 3D seismic reflection data, Quaternary Science Reviews, v. 26, p. 627–643.

Graham, A. G., Stoker, M. S., Lonergan, L., Bradwell, T., and Stewart, M. A., 2011, The Pleistocene Glaciations of the North Sea basin. Ehlers, J. and Gibbard, P.L. (eds.): Quaternary Glaciations 1 – Extent and Chronology, 2nd edition, p. 261–278

Gibbard, P.L., 1995, The formation of the Strait of Dover, Geological Society, London, Special Publications, v. 96, p. 15–26.

Gibbard, P.L., and Cohen, K. M., 2008, Global chronostratigraphical correlation table for the last 2.7 million years. Episodes, v. 31, p. 243–247.

Gibbard, P.L., and Cohen, K.M., 2015, Quaternary evolution of the North Sea and the English

Channel, Proceedings of the Open University Geological Society, v. 1, p. 63–74

Gibbard, P.L., and Lewin, J., 2016, Filling the North Sea Basin: Cenozoic sediment sources and river styles, Geologica Belgica, v. 19, p. 3–4.

Gupta, S., Collier, J. S., Palmer-Felgate, A., and Potter, G., 2007, Catastrophic flooding origin of shelf valley systems in the English Channel. Nature, v.448, p. 342-345.

Hamblin, R.J.O., Crosby, A., Balson, P.S., Jones, S.M., Chadwick, R.A., Penn, I.E. & Athur, M.J., 1992, United Kingdom offshore regional report: the geology of the English Channel. London: HMSO for the British Geological Survey

Henriet, J.P., De Moor, G., and De Batist, M. (eds.), 1989, The Quaternary and Tertiary Geology of the Southern Bight, North Sea, International Colloquy on the Quaternary and Tertiary.

Hijma, M. P., Cohen, K. M., Roebroeks, W., Westerhoff, W. E., and Busschers, F. S., 2012, Pleistocene Rhine–Thames landscapes: geological background for hominin occupation of the southern North Sea region. Journal of Quaternary Science, v. 27, p. 17–39.

James, J.W.C., Guennoc, P., Harrison, M., Le Bot, S., Philpott, S., Vinchon, C., Bee, E., Simien, F., Janjou, D., Garlan, T., Trentesaux, A., Mahieux, G., Briet D. and, Augris C., 2002, GEOSYNTH: A synthesis of the geology and sediments of the Dover Strait and its hinterland. CR/02/078, CDROM, British Geological Survey.

Lee, J. R., Busschers, F. S., and Sejrup, H. P., 2012, Pre-Weichselian Quaternary glaciations of the British Isles, The Netherlands, Norway and adjacent marine areas south of 68 N: implications for long-term ice sheet development in northern Europe, Quaternary Science Reviews, v. 44, p. 213–228.

Lericolais, G., Auffret, J.-P., and Bourillet, J.-F., 2003, The Quaternary Channel River: seismic stratigraphy of its palaeovalleys and deeps, Journal of Quaternary Science, v. 18, p. 245–260.

Lewis, S., Maddy D., and Glenday S., 2004, The Thames valley sediment conveyor: fluvial system development over the last two interglacial-glacial cycles. Quaternaire, v. 15, p. 17–28.

Liu, A. C., Missiaen, T., and Henriet, J. P., 1992, The morphology of the top-Tertiary erosion surface in the Belgian sector of the North Sea, Marine Geology, v.105, p. 275–284.

Mansy, J.-L., Manby, G.M., Averbuch, O., Everaerts, M., Bergerat, F., Van Vliet-Lanoë, B., Lamarche, J. & Vandycke, S., 2003, Dynamics and inversion of the Mesozoic Basin of the Weald-Boulonnais area: role of basement reactivation, Tectonophysics, v. 373, p. 161–179.

Mathys, M., 2009, The Quaternary geological evolution of the Belgian Continental Shelf, southern North Sea, PhD thesis, Ghent University, Belgium, 454 p.

Mellett, C.L., Mauz, B., Hodgson, D.M., Plater, A.J., and Lang, A., 2012, Optical dating of drowned landscapes: a case study from the English Channel. Quaternary Geochronology, v. 10, p. 201–208.

Mellett, C., Hodgson, M., Plater, A., Mauz, B., Selby, I. & Lang, A., 2013, Denudation of the continental shelf between Britain and France at the glacial–interglacial timescale, Geomorphology, v. 203, p. 79–96.

Minguely, B., Averbuch, O., Patin, M., Rolin, D., Hanot, F. & Bergerat, F., 2010, Inversion tectonics at the northern margin of the Paris basin (northern France): new evidence from seismic profiles and boreholes interpolation in the Artois area, Bulletin de la Société Géologique de France, v. 181, p. 429–442.

Oggioni, F., 2013, Characterization of catastrophic flood-related features in the English Channel. PhD thesis, Imperial College London, 330 p.

Ottemöller, L., Baptie, B., and Smith, N.J.P., 2009, Source Parameters for the 28 April 2007 Mw 4.0 Earthquake in Folkestone, United Kingdom, Bulletin of the Seismological Society of America, v. 99, p. 1853–1867.

Patton, H., Hubbard, A., Andreassen, K., Auriac, A., Whitehouse, P. L., Stroeven, A. P., Shackleton.C., Winsborrow, M., Heyman, J., and Hall, A. M., 2017, Deglaciation of the Eurasian ice sheet complex, Quaternary Science Reviews, v. 169, p. 148–172.

Peeters, J., Busschers, F. S., and Stouthamer, E., 2015, Fluvial evolution of the Rhine during the last interglacial-glacial cycle in the southern North Sea basin: a review and look forward, Quaternary International, v. 357, p. 176–188.

Roep, T. B., Holst, H., Vissers, R. L. M., Pagnier, H., and Postma, D., 1975, Deposits of southward-flowing, pleistocene rivers in the channel region, near Wissant, NW France. Palaeogeography, Palaeoclimatology, Palaeoecology, v. 17, p. 289–308.

Sejrup, H. P., Nygård, A., Hall, A. M., and Haflidason, H., 2009, Middle and Late Weichselian (Devensian) glaciation history of south-western Norway, North Sea and eastern UK, Quaternary Science Reviews, v. 28, p. 370-380.

Sejrup, H. P., Clark, C. D., and Hjelstuen, B. O., 2016, Rapid ice sheet retreat triggered by ice stream debuttressing: Evidence from the North Sea. Geology, v. 44, p. 355–358.

Smith, A.J., 1985, A catastrophic origin for the palaeovalley system of the eastern English Channel. Marine Geology, v. 64, p. 65–75.

Toucanne, S., Zaragosi, S., Bourillet, J. F., Gibbard, P. L., Eynaud, F., Giraudeau, J., Turon, J.L., Cremer, M., Cortijo, E., Martinez, P., and Rossignol, L., 2009a, A 1.2 Ma record of glaciation and fluvial discharge from the West European Atlantic margin. Quaternary Science Reviews, v. 28, p. 2974–2981.

Toucanne, S., Zaragosi, S., Bourillet, J.F., Cremer, M., Eynaud, F., Van Vliet-Lanoë, B., Penaud, A., Fontanier, C., Turon, J.L., Cortijo, E. and Gibbard, P.L., 2009b, Timing of massive 'Fleuve Manche' discharges over the last 350 kyr: insights into the European ice-sheet oscillations and the European drainage network from MIS 10 to 2, Quaternary Science Reviews, v. 28, p. 1238–1256.

Toucanne, S., Zaragosi, S., Bourillet, J. F., Marieu, V., Cremer, M., Kageyama, M., Van Vliet-Lanoë, B., Eynaud, F., Turon, J-L, and Gibbard, P. L., 2010, The first estimation of Fleuve Manche palaeoriver discharge during the last deglaciation: evidence for Fennoscandian ice sheet meltwater flow in the English Channel ca 20–18 ka ago, Earth and Planetary Science Letters, v. 290, p. 459–473.

Toucanne, S., Soulet, G., Freslon, N., Jacinto, R. S., Dennielou, B., Zaragosi, S., Eynaud, F., Bourillet, J-F., and Bayon, G., 2015, Millennial-scale fluctuations of the European Ice Sheet at the end of the last glacial, and their potential impact on global climate. Quaternary Science Reviews, v. 123, p. 113-133.

Van Vliet-Lanoë, B., Mansy, J.L., Margerel, J.P., Vidier, J.P., Lamarche, J., Everaerts, M., 1998, The Dover Strait, a discretely open Cenozoic strait, C.R. Accad. Sci. Paris, Earth & Planetary Sciences, v. 362, p. 729–736.

Van Vliet-Lanoë, B., Vandenberghe, N., Bergerat, F., Henriet, J.P., Mansy, J.L., Meilliez, F., Lacquement, F., Sintubin, M. and Vandycke, S., 2002a, Cenozoic evolution of the Channel and southern North Sea areas (Western Europe). The neotectonic control on sedimentation and rivers. In Contributions to the Geology of Belgium and Northwest Europe: Proceedings of the First Geologica Belgica International Meeting, Leuven, v. 12, p. 21-24.

Van Vliet-Lanoë, B., Vandenberghe, N., Laurent, M., Laignel, B., Lauriat-Rage, A., Louwye, S., Mansy, J.L., Mercier, D., Hallégouët, B., Laga, P. and Laquement, F., 2002b, Palaeogeographic evolution of northwestern Europe during the Upper Cenozoic, Geodiversitas, v.24, p.511–541.

Van Vliet-Lanoë, B., Mansy, J.L., Henriet, J.-P., Laurent, M. & Vidier, J.P., 2004. Une inversion tectonique cénozoïque par étapes : le Pas-de-Calais, Bulletin Société géologique Français, v.2, p. 175–195.

Westaway, R., & Bridgland, D. R., 2010, Causes, consequences and chronology of large-magnitude palaeoflows in Middle and Late Pleistocene river systems of northwest Europe. Earth Surface Processes and Landforms, v.35, p. 1071–1094.

Chapter 2

Geological Setting

Chapter 2– Geological Setting

The North Sea is a continental shelf sea located between northwestern continental Europe, the Scandinavian Peninsula and Great Britain (Figure 2.1). This sea enters the Atlantic Ocean through the Dover Strait/English Channel to the southwest, and via the Norwegian Sea (in between Great Britain and Norway) to the North (Figure 2.1). It is commonly sub-divided in three parts; i.e. the southern North Sea, the central North Sea and the northern North Sea (Figure 2.1).

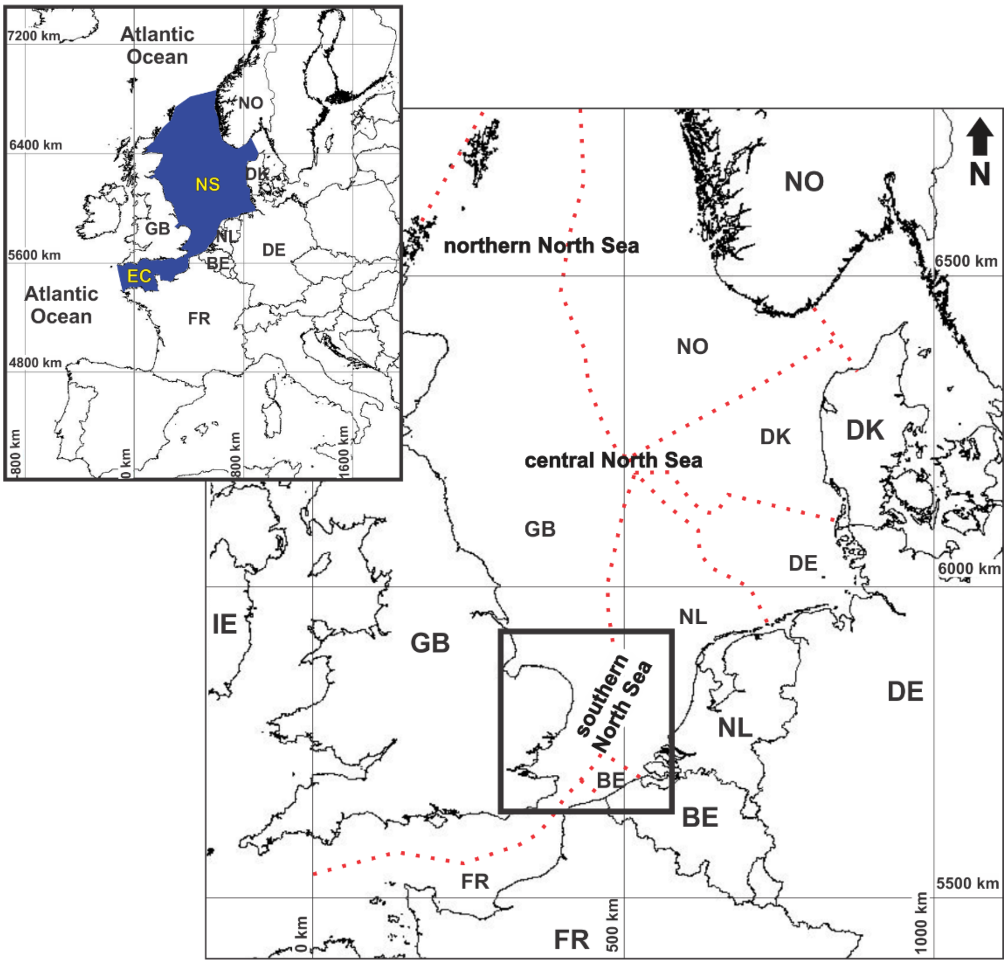

Figure 2.1. Geographical situation of the North Sea and English Channel. Offshore national boundaries are indicated (red-dotted lines). For this and next figures, projections: UTM zone 31N/WGS84; distances are given in kilometers.

The North Sea continental shelf occupies an area of ~575000 km², exhibiting average depths of 95 m. In the present study, we will investigate the geomorphology of the part of the southern North Sea that was not glaciated during the Last Glacial Maximum (Figure 1.1). Here below, we briefly summarize the geology of that area.

2.1 Pre-Quaternary geology

The basement of the southern North Sea was formed during the Precambrian and Lower Paleozoic Eons (Figure 2.2; e.g. Cameron et al., 1993). It mainly comprises marine and volcanoclastic Lower Paleozoic sediments, which were metamorphosed during the Caledonian orogeny (420–390 Ma ago). Crystalline Precambrian rocks are also present at depth beneath the London-Brabant Massif (e.g. Cameron et al., 1993). During the Devonian (410–360 Ma ago), there was widespread red-bed molasse and lacustrine sedimentation as the newly-formed Caledonian mountain ranges were eroded (e.g. Cameron et al., 1993). In the Middle Devonian (~375 Ma ago) a rift phase started, resulting in marine limestone deposition (Cameron et al., 1993).

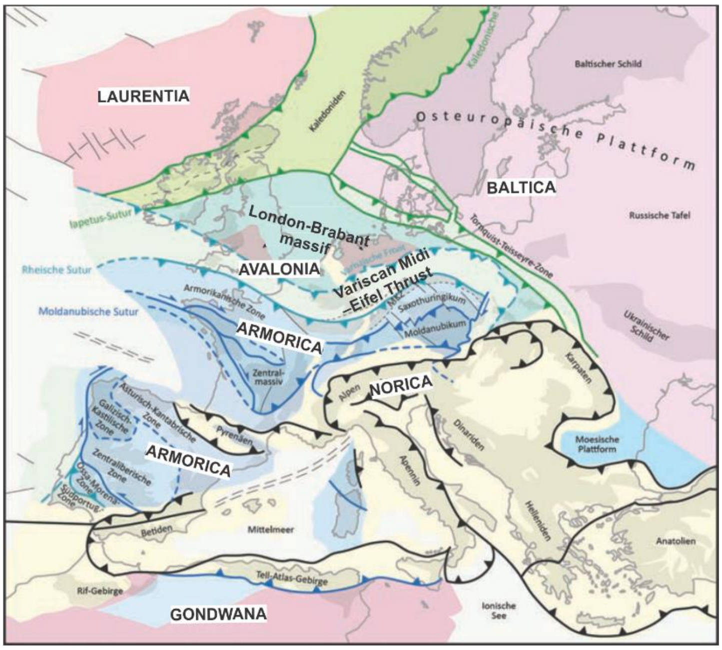

Figure 2.2. Tectonostratigraphic units of Europe. Pink areas: Precambrian basement; green areas: Caledonian Orogeny; blue areas: Variscan Orogeny; yellow areas: Alpine Orogeny. After Harff et al. (2017).

The London-Brabant Massif became an upland area early in Carboniferous times (e.g. Cameron et al., 1993). During that period, up to 4 km of deep-water and deltaic sediments were deposited in rapidly subsiding grabens due to crustal extension (e.g. Leeder, 1987). Carboniferous deposits were later gently folded, faulted, uplifted and eroded 300–290 Ma ago due to the Variscan Orogeny (Cameron et al., 1993).

During the Permian System (300–252 Ma), much of the southern North Sea began to subside again due to post-orogenic collapse (e.g. Ziegler, 1990). However, the London-Brabant Massif remained a stable upland area during much of this period (Cameron et al., 1993). Permian sediments include Lower Permian aeolian and fluvial sands and Upper Permian marine shales, carbonates and evaporites (e.g. Cameron et al., 1993).

The Mesozoic was characterized by rifting and basin inversion episodes (e.g. Cameron et al., 1993). It was during Late Triassic–Early Jurassic crustal extension that the North Sea first adopted its present NNW-SSE alignment (e.g. Cameron et al., 1993). During that time, Lower Triassic sandstones, mudstones and evaporites were overlain by Lower Jurassic argillaceous marine sediments. The subsidence was interrupted in the Middle Jurassic due to an uplift episode (e.g. Cameron et al., 1993). This uplift was however inverted in Late Middle Jurassic times by an extensional episode that lasted until the Middle Cretaceous Epoch. The subsidence induced during that period enabled the deposition of several kilometers of Upper Jurassic and Lower Cretaceous marine shales, limestone, calcareous mudstones and sandstones. Crustal extension ended in the Middle Cretaceous Epoch, but the regional thermal anomaly that it generated continued decaying, inducing subsidence. Notably, a thick (up to 1500 m) layer of chalk accumulated in Late Cretaceous times.

Chalk deposition was interrupted in the Paleogene by tectonic uplift generated by the continental convergence during the Alpine Orogeny (Figure 2.2). That caused a basin inversion, which induced basin uplift and temporary emergence of the southern North Sea, resulting in an erosional phase. That stress system relaxed in the Late Miocene (~10 Ma ago), enabling subsidence to resume. At that time and until the Middle Pleistocene (~400 ka BP), major delta systems developed across the southern North Sea Basin, which resulted in the deposition of several hundreds-of-meters thick delta-related sedimentary formations.

The Paleozoic basement is thus buried beneath many kilometers of younger Paleozoic, Mesozoic and Cenozoic sediments. Pre-Quaternary formations outcropping or sub-cropping beneath the Quaternary in the North Sea are mainly composed of locally faulted and gently folded Paleogene and Neogene clays, sands, mudstones and sandstones (Figure 2.3; e.g. Cameron et al., 1984a; Balson et al., 1991; Le Bot et al., 2003).

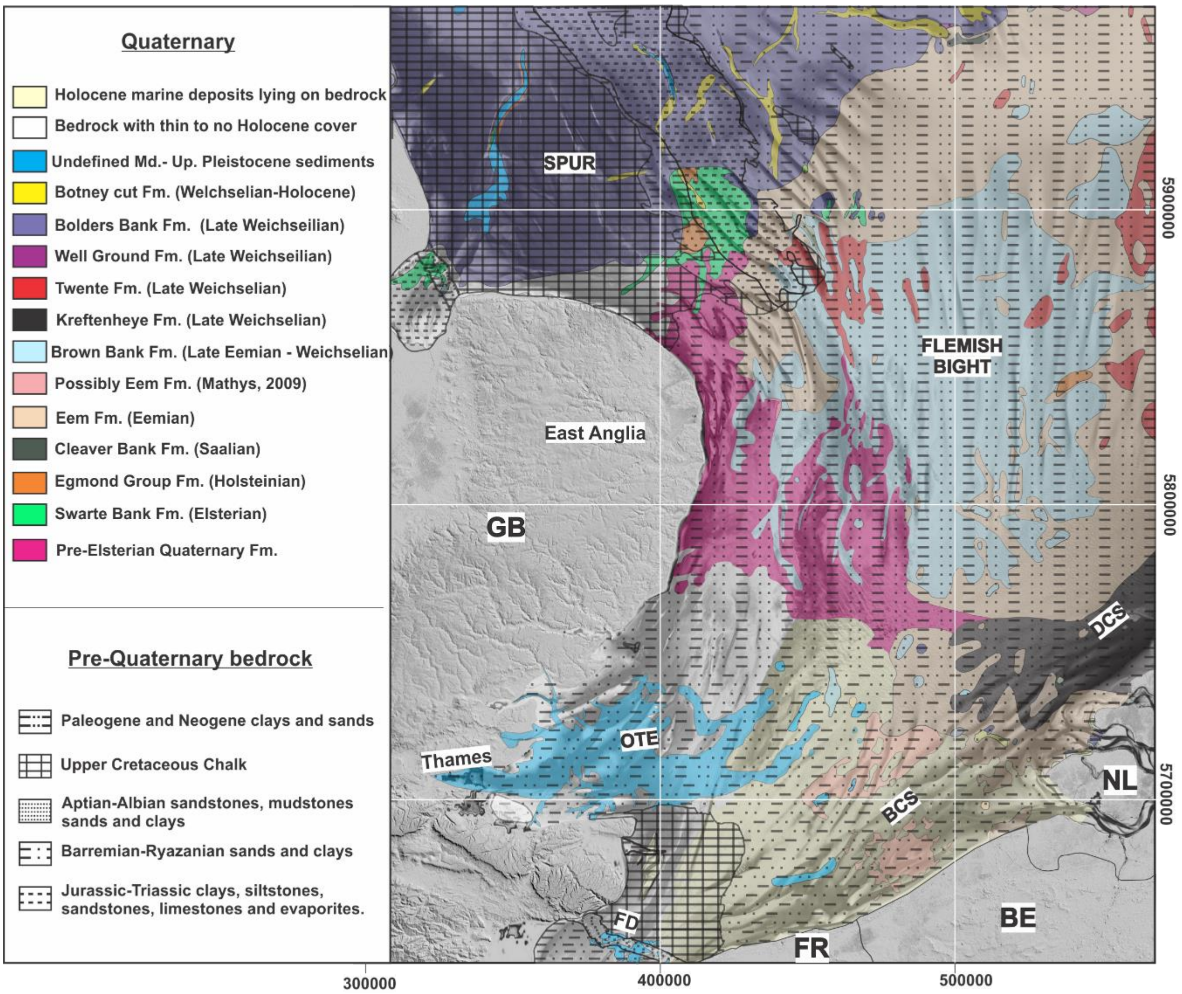

Figure 2.3. Geological map of the southern North Sea. Modified from British Geological Survey (BGS) maps produced by Cameron et al. (1984a), Cameron et al. (1984b), Gaunt et al. (1985), Zalasiewicz and Balson (1985), Cameron et al. (1986); Cameron et al. (1987), Balson and D'Olier (1989), Balson and D'Olier (1990), Balson et al. (1991), Tappin (1991), and Laban et al. (1992). FD: Fosses Dangeard; BCS: Belgian Continental Shelf; DCS: Dutch Continental Shelf; OTE: Outer Thames Estuary.

The oldest sedimentary formations outcropping in the southern North Sea are Jurassic and Cretaceous chalk, clays, mudstones, sandstones, limestones and sands (Figure 2.3). These formations outcrop in the Dover Strait and in the Spur area. In the Dover Strait, they are intensively deformed by the Variscan Weald–Artois anticline and the North Artois Shear Zone (Figure 2.4). As mentioned above, these tectonic structures passed through a series of reactivations and tectonic inversions since their formation until the Miocene Epoch. However, their activity since then is unknown. It has been suggested though that the Weald-Artois Shear Zone may still be accommodating some tectonic deformation (Colbeaux et al., 1981; Van Vliet-Lanoë et al., 2002; Mansy et al., 2003).

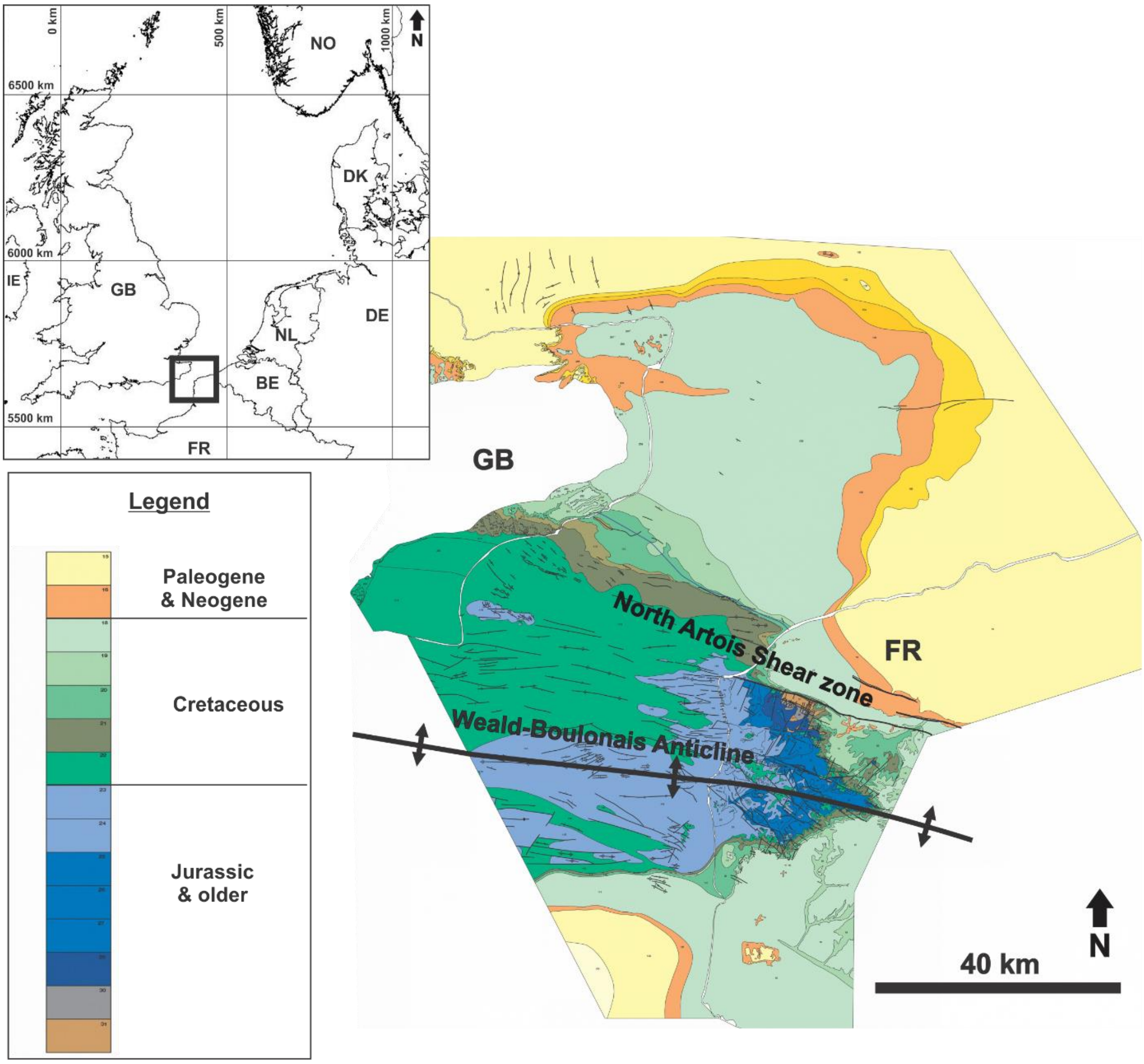

Figure 2.4. Dover Strait bedrock geology showing major tectonic structures; modified from James et al. (2002).

2.2 Quaternary geology

2.2.1 Pleistocene

The Pleistocene Epoch comprised several extreme climatic changes, resulting in a succession of interglacial and glacial stages that involved major changes in sea level and several episodes of glacial isostatic depression and post-glacial isostatic rebound (e.g. Zagwijn, 1989; Kjemperud and Fjeldskaar, 1992; Ehlers and Gibbard, 2004; Busschers et al., 2007). Each of these climatic changes significantly modified the location, extent, morphology and dynamics of the different northwestern European drainage systems (e.g. Gibbard, 1995; Lewis et al., 2004; Busschers et al., 2005; Busschers et al., 2007; Hijma et al., 2012; Peeters et al., 2015;

Peeters et al., 2016; Gibbard and Cohen, 2015). The present-day geography and geomorphology of the southern North Sea and English Channel are therefore the result of a polycyclic process.

Here below, we provide brief descriptions of the main Pleistocene glacial and interglacial periods that affected the North Sea during the time interval targeted in the present study; i.e. between the supposed age of the opening of the Dover Strait (~450 ka BP) and the present. In this section, we mainly focus on the regional palaeogeographic/hydrographic changes that occurred in the southern North Sea – English Channel during the glacial maxima of the various Middle–Late Pleistocene glaciations, during which the erosional features investigated in this dissertation appear to have been formed.

Around 450 ka ago (MIS 12), Earth was in the middle of a major glacial stage known as the Elsterian glacial maximum. This glaciation induced the widespread extension of ice across the North Sea Basin and a major sea level drop. During this glacial stage, the Irish–British ice sheets reached their maximum extent of the Quaternary Period, coalescing for the first time with the Scandinavian ones across the North Sea (e.g. Ehlers and Gibbard, 2004). This resulted in the incision of hundreds-of-meters deep, kilometer-scale glacial valleys in the areas of the North Sea under the ice (e.g. Kristensen et al., 2007; Sejrup et al., 2009). In this palaeogeographic context, the unglaciated part of the southern North Sea and the English Channel were emerged (e.g. Gibbard, 1995). More importantly, the merged ice sheet blocked the northern drainage routes of southern North Sea palaeo-rivers (Figure 2.5; e.g. Gibbard, 1995). This, added to the subsidence induced by glacial isostasy, seems to have resulted in the formation of a major lake in the southern North Sea (Figure 2.5; e.g. Gibbard, 1995; Murton and Murton, 2012; Gibbard and Cohen, 2015; Gibbard and Lewin, 2016). Evidence for such a lake comes from glacio-lacustrine deposits identified at several locations along the coast of East Anglia and in the North Sea (Ter Wee 1983; Banham, 1988; Hart, 1992; Lunkka, 1994; Gibbard, 1995). That lake was purportedly dammed in the southwest by the Weald-Artois ridge, which extended across the Dover Strait (e.g. Gibbard and Lewin, 2016). The dimensions of that lake are uncertain. It has been speculated that it might have covered an area of ~140,000 km², extending several tens of kilometers into the present-day lands of the Netherlands and Belgium (Gibbard et al., 1988; Gibbard and Clark, 2011, Ehlers and Gibbard, 2007; Murton and Murton, 2012; Gibbard and Cohen, 2015). The possible inundation of such a large area implies that the Dover Strait had higher elevation than the inundated parts of Belgium and the Netherlands. That is explained by the presence of the Weald–Artois ridge traversing the Dover Strait and the possible greater subsidence along the ice-sheet southern margin than in the Dover Strait (Gibbard and Cohen, 2015). This lake seems to have outflowed through the then closed Dover Strait, overspilling the dam formed at that location by the Weald–Artois ridge (Smith, 1985; Ehlers and Rose, 1991; Gibbard, 1995; Gibbard and Cohen, 2015). Several authors have hypothesized that the Weald–Artois ridge breached at some point during the Elsterian glacial maximum, thus initiating the opening of the Dover Strait (e.g. Gibbard, 1988; Gibbard, 1995; Gupta et al., 2007; Gibbard and Cohen, 2015). That is consistent with sediment cores collected from the Gulf of Biscay, which suggest the occurrence of massive fluvial discharges from the Channel palaeo-river during that glacial stage (Toucanne et al., 2009a). Whether the breach happened due to progressive fluvial erosion or due to a sudden catastrophic breach of the ridge is currently unknown.

The Elsterian glaciation was followed by the Holsteinian interglacial (Meijer and Preece, 1995; Gibbard, 1995; Gibbard et al., 2007); a warm period that took place between 424–374 ka BP (MIS 11). During this Interglacial, most of the Elsterian continental ice sheets melted and sea level rose, submerging the English Channel and southern North Sea. Studies of fossil mollusks and pollen indicate that the climatic conditions prevailing then were similar to those of today (e.g. Meijer and Preece, 1995). Biostratigraphic and sedimentary data also indicate that the land-bridge connecting Great Britain and continental Europe may still have been in place during most of this interglacial (e.g. White and Schreve, 2000). It appears that it was not until the Holsteinian highstand maximum, that marine exchange between these basins started (Meijer and Preece, 1995; Gibbard, 1995; White and Schreve, 2000). It is therefore likely that the breach opened in the Weald-Artois ridge during the previous glaciation was significantly widened during this interglacial period by tidal and coastal erosion.

The climatic conditions deteriorated again ~374 ka BP (MIS 10), giving place to the Saalian glaciation that lasted until 130 ka BP (MIS 6). This glaciation had two glacial maxima, i.e. the Drenthe and the Warthe glacial advances, which took place during MIS 6, at 175–155 ka BP and 150–140 ka BP, respectively (e.g. Toucanne et al., 2009b, Gibbard and Cohen, 2015). The Scandinavian and Irish–British ice sheets appear to have merged again across the North Sea during both glacial maxima, closing once more the northern drainage routes of southern North

Sea palaeo-rivers (Ehlers and Gibbard, 2004). This, added to the isostatic depression induced by the weight of the glacier, appears to have resulted in the formation of another ice-marginal lake in the southern North Sea (e.g. Gibbard and Cohen, 2015). Sedimentological and biostratigraphic data indicate that this lake extended tens of kilometers into The Netherlands (e.g. Meijer and Preece, 1995; Busschers et al., 2008). However, it did not extend over the Belgian western coastal plain (e.g. Bogemans et al., 2016). This is consistent with recent palaeogeographic reconstructions, which proposed that the Saalian southern North Sea lake was confined in between the southern margin of the ice sheet and a land-bridge that extended across the southern North Sea from southern East Anglia to the border between Belgium and The Netherlands (Figure 2.5). The nature of that land-bridge is currently unknown. Gibbard and Cohen (2015) hypothesized that it may have been composed of former glacial moraines. That hydrographic configuration appears to have evolved into a continuous major river system around 150–140 ka BP (still within MIS 6), comprising the Meuse, the Rhine, the Thames and the Channel Palaeo-river (Figure 2.5; see also Gibbard, 1995; Gibbard and Cohen, 2015). It is uncertain whether this happened because of a sudden breach of the land-bridge or due to progressive fluvial erosion. Nevertheless, sedimentological data indicate the occurrence of massive fluvial discharges in the Gulf of Biscay at the Drenthe glacial maximum (Toucanne et al., 2009b). It is thus possible that emptying of the Saalian lake was also related to a sudden (catastrophic) breach of the land-bridge damming it in the south, which would have induced intense erosion in the Dover Strait and English Channel (see also Gupta et al., 2007; Collier et al., 2015; Gibbard and Cohen, 2015).

The Saalian glaciation ended around 130 ka, when the southern North Sea was submerged again due to warmer climatic conditions. This period is known in Northwestern Europe as the Eemian interglacial (MIS 5e; 130–110 ka). The thermal maximum of this stage appears to have been warmer than the Holocene one, resulting in slightly higher sea levels than at the present (e.g. Turner, 2000). The Dover Strait was therefore subjected once again to tidal and coastal erosion during this highstand, which may have contributed to the widening of the Strait. This period involved significant deposition of marine and deltaic sediments in the Flemish Bight (see Figure 2.3).

The sea retreated again during the Weichselian glaciation, also known as Last Glacial Age (MIS 5d–2), which lasted until 12 ka ago (e.g. Gibbard et al., 2007). During this glacial period, the

British–Irish and Scandinavian ice sheets appear to have merged again across the North Sea at 30–19 ka BP, and, perhaps, at ~70 ka BP (e.g. Carr et al., 2006). This resulted in the deviation of palaeo-rivers running into the unglaciated part of the southern North Sea, which started heading toward the Dover Strait (Figure 2.5; e.g. Gibbard et al., 2007; Toucanne et al., 2010; Hijma et al., 2012; Toucanne et al., 2015). At these stages, the Rhine–Meuse and Thames palaeo-rivers converged again in the Axial Channel, which joined the Channel River through the Dover Strait along the Lobourg Channel (Figure 1.1; e.g. Gibbard, 1995; Hijma et al., 2012). This palaeo-river system appears to have had significant fluvial input from the north, either due to a major river formed by several southward diverted German rivers (Figure 1.1; Toucanne et al., 2010; Toucanne et al., 2015) or from the main outflow of a major ice-marginal lake located on the German–Danish continental shelf, into which northern German rivers discharged (Figure 2.5; Sejrup et al., 2016; Patton et al., 2017).

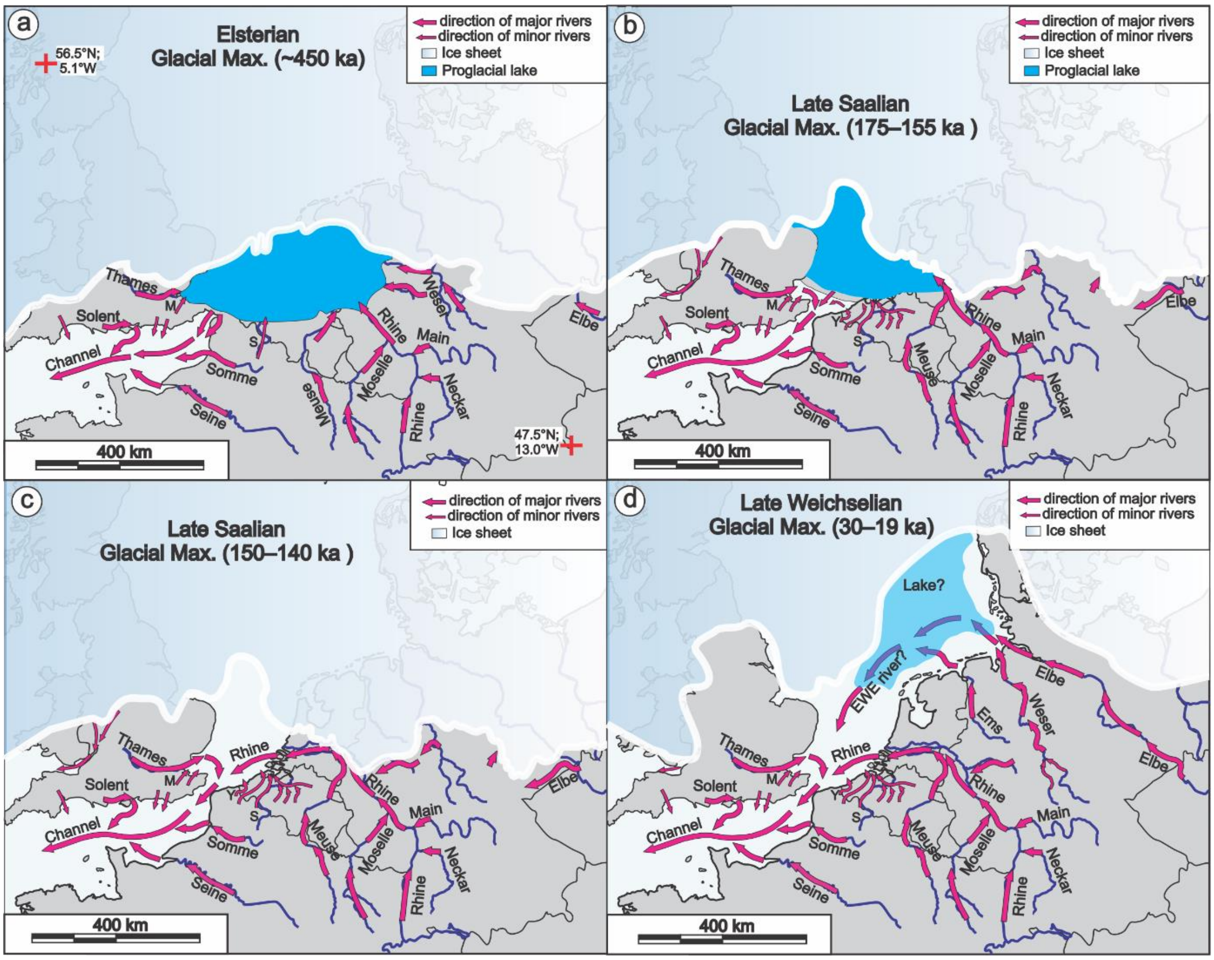

Figure 2.5. Paleo-geographic reconstructions major Middle–Late Pleistocene glacial maxima. a) Elsterian glacial maximum. Modified from Cohen et al. (2014), Gibbard and Cohen (2015), and Gibbard and Lewin (2016). b) Drenthe glacial Maximum, after Gibbard and Cohen (2015). c)

Warthe glacial maximum (Gibbard, 1988; Gibbard, 1995); d) Weichselian glacial maximum (Gibbard, 1988; Gibbard, 1995; Sejrup et al., 2016). Possible Last Glacial ice-marginal lake is indicated (Sejrup et al., 2016). Possible flow directions of the Ems, Weser and Elbe Rivers across the southern North Sea (EWE River) during the Last Glacial Maximum are also indicated (after Toucanne et al., 2015). Y: Yser River; S: Scheldt River; pink arrows: flow direction of main drainage systems.

These changes in climatic conditions resulted in a succession of subaerial and submarine erosional/depositional episodes, which generated the incision of palaeovalleys and palaeo-depressions in the southern North Sea and English Channel, as well as the deposition of a series of marine, glacial and fluvial Middle–Late Pleistocene sedimentary formations in the North Sea. Apart from the infills of the Fosses Dangeard and some Holocene sand ridges and dunes, few sediments remain from those episodes in the Dover Strait. The southern North Sea, on the other hand, comprises a series of marine, glacial and fluvial Middle–Late Pleistocene sedimentary formations outcropping at the seafloor or sub-cropping beneath the Holocene cover. Here below, we describe briefly the main Pleistocene sedimentary formations found in that area (Figure 2.4).

Pre-Elsterian Quaternary formations include marine, locally shelly, clays and sands, which evolve upwards into fluvial sands with clay intercalations (Cameron et al., 1984a). These units reach locally hundreds of meters in thickness (Cameron et al., 1984a).

The Swarte Bank Formation refers to sediments deposited during the Elsterian glacial period (MIS 12; 480–400 ka BP), which infill a series of kilometric-scale buried palaeo-valleys several hundreds of meters deep (Cameron et al., 1986). This formation includes gravelly sands, clays and silts deposited in glacial, glacio-lacustrine and fluvioglacial environments (e.g. Cameron et al., 1986).

The Egmond Ground Formation was deposited during the Holsteinian interglacial (possibly at MIS 11; 400–350 ka BP). This sedimentary formation comprises sparsely shelly marine sands interbedded by silt and clay bands that can reach tens of meters in thickness (e.g. Cameron et al., 1986). The Edmond Ground and the Swarte Bank formations outcrop locally at the seafloor in the Spur area; elsewhere, they sub-crop beneath younger Pleistocene and/or Holocene sedimentary formations (e.g. Cameron et al., 1986).

The Saalian (350–130 ka BP) Cleaver Bank Formation sub-crops locally beneath the Holocene sands in the northern part of the study area. This sedimentary formation is ~10 m thick and comprises silty clays with silt and sand laminae, and fluvioglacial sands interbedded with silt and clay (e.g. Cameron et al., 1986).

The Eemian (130–110 ka BP) Eem Formation extends throughout most of the eastern half of the study area, mostly sub-cropping beneath younger Pleistocene and Holocene sediments. This formation is composed of gravelly and shelly marine sands that evolve upward to muddy sands (e.g. Cameron et al., 1986). In the southern North Sea, the Eem Formation is 5–20 m thick (e.g. Cameron et al., 1986).

The Weichselian (110–12 ka BP) and pre-marine transgression Holocene sedimentary formations (i.e. younger than 7000–6000 years BP; see Sturt et al., 2013) found in the southern North Sea are: the Brown Bank Formation, the Kreftenheye Formation, the Well Ground Formation, the Bolders Bank Formation, the Twente Formation, and the Botney Cut Formation (e.g. Cameron et al., 1984b; Cameron et al., 1986).

The Brown Bank Formation was deposited during the Eemian–early Weichselian marine regression (e.g. Cameron et al., 1986). It extends over large areas of the central part of the southern North Sea, outcropping locally near the coastline of East Anglia. This sedimentary formation is up to 5 m thick, and it is composed of silty clay with silt and sands interbedded with shelly gravelly sands toward the base (e.g. Cameron et al., 1986).

The Kreftenheye Formation is mostly located on the Dutch continental shelf. This stratigraphic unit comprises sands, gravels and reworked Eemian shells deposited by the Rhine–Meuse Palaeo-river system during the Weichselian glacial period (Cameron et al., 1984b).

The Well Ground Formation comprises Late Weichselian fluvioglacial sands, silts and clays and has a maximum thickness of 5 m (Cameron et al., 1986).

The Bolders Bank Formation is composed of gravelly and sandy clay till. This unit extends over most of the northwestern part of the study area. It is associated with Weichselian glacial erosional/depositional processes, and it has a maximum thickness of 15 m (Cameron et al., 1986).

The Twente Formation comprises sands that were deposited during Late Weichselian – Early Holocene stages (Cameron et al., 1986). This sedimentary formation is 1–5 m thick and sub-crops locally beneath younger Holocene sediments in the eastern and central parts of the study area.

Finally, the Botney Cut Formation refers to the infill of Weichselian kilometric-scale glacial valleys, which locally reach depths of 80 m. It includes gravelly sands and glacio-lacustrine sandy muds (Cameron et al., 1986).

2.2.2 Holocene

The Holocene Epoch is characterized by warm climatic conditions. This climatic setting induced a progressive regional transgression 12000 years ago, which reached the present-day marine level at 7000–6000 years BP (Figure 2.6; e.g. Sturt et al., 2013).

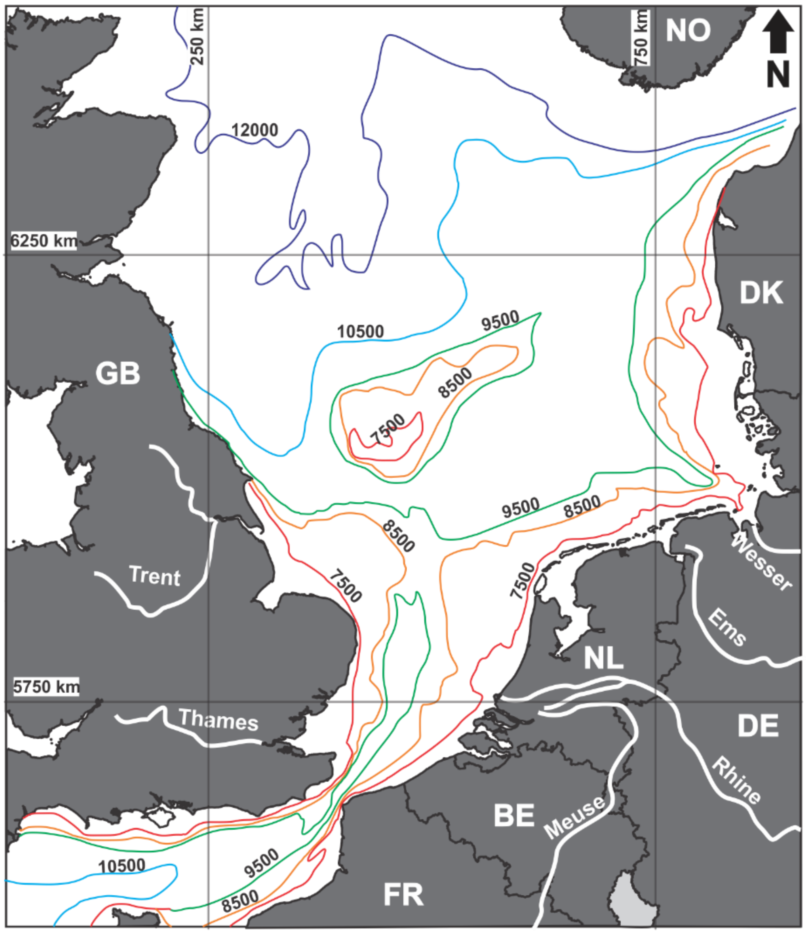

Figure 2.6. Different positions of shorelines during the Holocene marine transgression that took place between 12,000 and 6,000 years BC. Modified from Sturt et al. (2013).

Currently, the southern North Sea and the Dover Strait are subjected to large differences between low and high tides, which reach up to 7 meters in the Dover Strait (Figure 2.7). This results in strong submarine currents. The marine transgression and the establishment of the present-day submarine currents during the Holocene Epoch have resulted in further seafloor erosion and sediment remobilization in the Dover Strait (e.g. Reynaud et al., 2003; Hamblin et al., 1992), as well as in continuous coastal erosion of the cliffs at its edges. The marine transgression also induced the formation of a series of kilometer-scale, elongated sandbanks, megaripples and dunes in the southern North Sea and Dover Strait (Houbolt, 1968; Caston and Stride, 1973; Dyer and Huntley, 1999; Reynaud et al., 2003). Sandbanks (in this study also referred to as tidal sand ridges) appear to have been originally formed by the interaction of tidal currents, coastal recession and the seafloor palaeo-topography during the Holocene marine transgression (e.g. Caston and Stride, 1973; Dyer and Huntley, 1999). These features are in constant evolution due to the re-distribution of sand by submarine currents (e.g. Dyer and Huntley, 1999). The dunes and ripples, on the other hand, are mobile sedimentary bodies formed by the action of submarine currents. Dunes and ripples are mainly developed orthogonally to sandbanks (e.g. Dyer and Huntley, 1999). The sediments composing the sandbanks and dunes are sourced by reworking of former littoral barriers and/or valley infills (e.g. Reynaud et al., 2003).

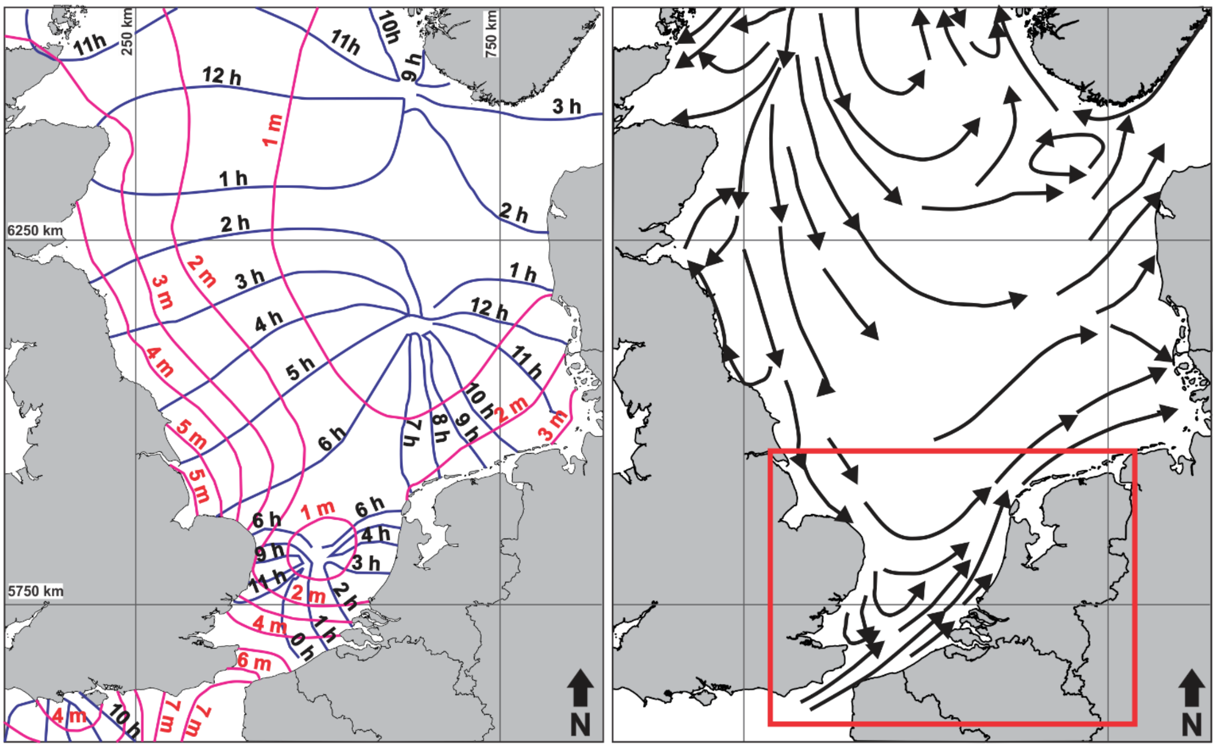

Figure 2.7. Present-day tidal-current systems in the North Sea. Left, lines of equal mean high-water in hours (blue lines) and Mean spring-tidal range in meters after the moon's transit over the meridian of Greenwich (red lines). Right, average surface current vectors. Modified from Houbolt (1968). Red box: Figure 2.8.

Sandbanks are mostly clustered in more or less extensive fields in the southern North Sea;

however, in the Dover Strait and English Channel, sand ridges are isolated (Figure 2.8; Dyer and Huntley, 1999). The main sandbank fields located in the southern North Sea are: the Norfolk Banks, the Flemish Bight Banks, the Belgian–Dutch Continental Shelf Banks and the Thames Estuary Banks. The tidal variation during the Holocene transgression (Figure 2.6; e.g. Shennan et al. 2000), and the present-day submarine current vectors across the North Sea and English Channel (Figure 2.7) have favored deposition in some areas and erosion in others. For instance, the south-westernmost area of the southern North Sea and the Dover Strait have almost no Quaternary sediments overlying the Cretaceous–Jurassic bedrock (e.g. Hamblin et al., 1992; James et al., 2002), while the Belgian Continental Shelf and the Flemish Bight are covered by extensive sandbank and dune fields (Dyer and Huntley, 1999).

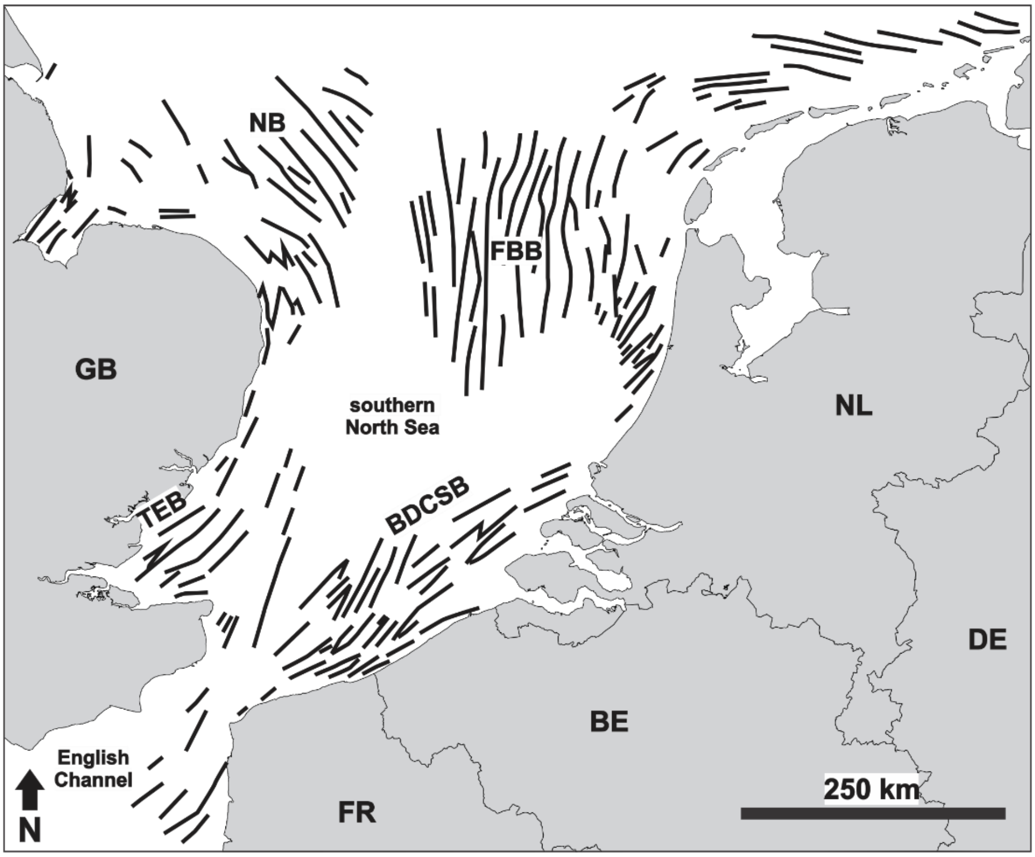

Figure 2.8. Location and schematic plan-view morphology of elongated sandbanks. NB: Norfolk Banks; FBB: Flemish Bight Banks; BDCSB: Belgian–Dutch Continental Shelf Banks; TEB: Thames Estuary Banks. Modified from Dyer and Huntley (1999).

References

Banham, P.H., 1988, Polyphase glaciotectonic deformation in the contorted drift of Norfolk. In D.G. Croor (ed), Glaciotectonics: forms and processes, 27–32. Rotterdam: Balkema

Bateman, M.D., Buckland, P.C., Chase, B., Frederick, C.D., and Gaunt, G.D., 2008. The late-Devensian proglacial Lake Humber: new evidence from littoral deposits at ferrybridge, Yorkshire, England, Boreas, v. 37, p. 195–210.

Bogemans, F., Baetaman, C., 2016, Incised Pleistocene valleys in the Western Belgium coastal plain: Age, origins and implications for the evolution of the Southern North Sea Basin. Palaeogeography, Palaeoclimatology, Palaeoecology, p. 456, v. 46–59.

Busschers, F. S., Weerts, H. J. T., Wallinga, J., Cleveringa, P., Kasse, C., De Wolf, H., and Cohen, K. M., 2005, Sedimentary architecture and optical dating of Middle and Late Pleistocene Rhine-Meuse deposits-fluvial response to climate change, sea-level fluctuation and glaciation, Netherlands Journal of Geosciences, v. 84, p. 25–41.

Busschers, F. S., Kasse, C., Van Balen, R. T., Vandenberghe, J., Cohen, K. M., Weerts, H. J. T., Wallinga, J., Johns, C., Cleveringa, P., and Bunnik, F. P. M., 2007, Late Pleistocene evolution of the Rhine-Meuse system in the southern North Sea basin: imprints of climate change, sea-level oscillation and glacio-isostacy, Quaternary Science Reviews, v. 26, p. 3216–3248

Busschers, F. S., Van Balen, R. T., Cohen, K. M., Kasse, C., Weerts, H. J., Wallinga, J., and Bunnik, F. P., 2008, Response of the Rhine–Meuse fluvial system to Saalian ice-sheet dynamics. Boreas, v. 37, p. 377–398.

Balson, P.S., and D'Olier, B.D., 1989, Thames estuary Sheet 51°N–00°, solid geology, British Geological Survey, 1:250 000 series, 1 sheet.

Balson, P.S., and D'Olier, B.D., 1990, Thames estuary sheet 51°N–00°, seabed sediments and Quaternary geology, British Geological Survey, 1:250 000 series, 1 sheet.

Balson PS, Laban C, Frantsen PJ, Parker, N., Henriet, J.P., De Batist and M., Mostaert, F, 1991. Ostend Sheet 51°N–02°E, Solid geology, British Geological Survey, 1:250.000 series, 1 sheet.

Cameron, T.D.J.; Grimshaw, S., Hall, D.M., Laban, C., Mesdag, C.S., Parker, N., and Schüttenhelm, R.T.E., 1984a, Flemish Bight sheet 52N–02E, Solid Geology, 1:250 000 Series, British Geological Survey, 1 sheet.

Cameron, T.D.J.; Grimshaw, S., Hall, D.M., Laban, C., Mesdag, C.S., Parker, N., and Schüttenhelm, R.T.E., 1984b, Flemish Bight sheet 52N–02E, Solid Geology, 1:250 000 Series, British Geological Survey, 1 sheet.

Cameron, T.D.J., Laban, C., Mesdag, C.S., and Schüttenhelm, R.T.E., 1986, Indefatigable sheet 53N–02E, Quaternary Geology, 1:250 000 Series, British Geological Survey, 1 sheet.

Cameron, T.D.J., Bulat, J., Parker, N., Frantsen, P., Laban, C., and Schüttenhelm, R.T.E., 1987, Indefatigable sheet 53N–02E, Solid Geology, 1:250 000 Series, British Geological Survey, 1 sheet.

Cameron, D., Van Doom, D., Laban, C., and Streit, H.J., 1993, Geology of the Southern North Sea Basin, In: Roeland, H., and Verhagen, H.J. (ed.) "Coastlines of the southern North Sea.", Proceedings, 8th Symposium on Coastal and Ocean Management, New Orleans, Louisiana, p. 14–26.

Carr, S. J., Holmes, R. V. D., Van der Meer, J. J. M., & Rose, J., 2006, The Last Glacial Maximum in the North Sea Basin: micromorphological evidence of extensive glaciation, Journal of Quaternary Science, v. 21, p. 131–153.

Caston, V. N. D., and Stride, A. H., 1973, Influence of older relief on the location of sand waves in a part of the southern North Sea, Estuarine and Coastal Marine Science, v.1, p. 379–238.

Colbeaux, J.P., Somme, J.&Tuffreau, A., 1981, Tectonique quaternaire dans le Nord de la France: l'apport du gisement paléolithique de Biache-Saint-Vaast, Bulletin de l'Association française pour l'étude du quaternaire, 18, 183–192.

Dyer, K.R., and Huntley, D.A., 1999, The origin, classification and modelling of sand banks and ridges, Continental Shelf Research, v.19 1285–1330.

Ehlers, J., and Gibbard, P. L. (Eds.), 2004, Quaternary glaciations-extent and chronology: part I: Europe. Developments in Quaternary Science, v. 2, Elsevier, Amsterdam.

Ehlers, J., and Gibbard, P.L., 2007, The extent and chronology of Cenozoic Global Glaciation. Quaternary International, v. 165, p. 6–20.

Ehlers, J. and Rose, J., 1991. Glacial deposits of Britain and Europe: General overview. In: J. Ehlers, P. Gibbard and J. Rose (Eds): Glacial deposits of Britain and Ireland. Balkema, Rotterdam, p. 493-501.

Gaunt, G.D., Crosby, A., and Hall, D.M., 1985, Spurn solid geology, Sheet 53°N–00°, British Geological Survey, scale 1:250 000, 1 sheet.

Gibbard, P.L., 1988, The history of the great northwest European nvers during the past three million years. Philosophical Transactions of the Royal Socity B, v. 318, p. 559–602.

Gibbard, P.L., 1995, The formation of the Strait of Dover, Geological Society, London, Special Publications, v. 96, 15–26.

Gibbard, P.L., Cohen, K.M., Boreham, S. & Moscariello, A., 2007, Global chronostratigraphical correlation table for the last 2.7 million years. Subcommission on Quaternary Stratigraphy, International Commission on Stratigraphy, Cambridge.

Gibbard, P.L., and Cohen, K.M., 2015, Quaternary evolution of the North Sea and the English Channel, Proceedings of the Open University Geological Society, v. 1, p. 63–74.

Gibbard, P.L., and Lewin, J., 2016, Filling the North Sea Basin: Cenozoic sediment sources and river styles, Geologica Belgica, v. 19, p. 3–4.

Grünthal, G., and Wahlström, R., 2012, The European-Mediterranean earthquake catalogue (EMEC) for the last millennium. Journal of Seismology, v. 16, p. 535-570.

Gupta, S., Collier, J. S., Palmer-Felgate, A., and Potter, G., 2007, Catastrophic flooding origin of shelf valley systems in the English Channel. Nature, v.448, p. 342-345

Hamblin, R.J.O., Crosby, A., Balson, P.S., Jones, S.M., Chadwick, R.A., Penn, I.E., and Athur, M.J., 1992, United Kingdom offshore regional report: the geology of the English Channel, London: HMSO for the British Geological Survey.

Harff, J., Flemming, N.C., Groh, A., Hünicke, B., Lericolais, G., Meschede, M., Rosentau, A., Sakellariou, D., Uscinowicz, S., Zhang, W., and Zorita, E., 2017, Sea Level and Climate, in: Submerged landscapes of the European Continental Shelf. Quaternary Paleoenvironments; Eds: Flemming, N.C., Harff, J., Moura, D., Burgess, A., and Bailey, G. John Wiley & Sons, 517 p.

Hart, .J.K., 1992, Proglacial glaciotectonic deformation and the origin of the Cromer Ridge push moraine complex north Norfolk, UK; *Boreas v.* 19, p. 165–180

Hijma, M. P., Cohen, K. M., Roebroeks, W., Westerhoff, W. E., and Busschers, F. S., 2012, Pleistocene Rhine–Thames landscapes: geological background for hominin occupation of the southern North Sea region, Journal of Quaternary Science, v. 27, p. 17–39.

Houbolt, J.J.H.C., 1968, Recent sediments in the southern bight of the North Sea, Geologie en mijnbouw, v. 47, p. 245–273.

Huyghe, P. and Mugnier, J.L., 1995. A comparison of inverted basins of the southern North Sea and inverted structures of the external Alps. Geological Society, London, Special Publications, 88(1), pp.339-353.

James, J.W.C., Guennoc, P., Harrison, M., Le Bot, S., Philpott, S., Vinchon, C., Bee, E., Simien, F., Janjou, D., Garlan, T., Trentesaux, A., Mahieux, G., Briet D. and, Augris C., 2002, GEOSYNTH: A synthesis of the geology and sediments of the Dover Strait and its hinterland. CR/02/078, CDROM, British Geological Survey.

Kjemperud, A. T., and Fjeldskaar, W., 1992, Pleistocene glacial isostasy—implications for petroleum geology. Tectonic modelling and its application to petroleum geology. Norwegian Petroleum Society, Special Publication, v. 1, p. 187-195

Kristensen, T. B., Huuse, M., Piotrowski, J. A., and Clausen, O. R., 2007, A morphometric analysis of tunnel valleys in the eastern North Sea based on 3D seismic data, Journal of Quaternary Science, v. 22, p. 801–815.

Laban, C., Schüttenhelm, R.T.E., Balson, P.S., Baeteman, C., Paepe, R., 1992. Ostend Sheet 51N–02E. 1:250 000 Series. Quaternary Geology. British Geological Survey – Rijks Geologische Dienst – Belgische Geologische Dienst. Nottingham – Haarlem – Brussel.

Le Bot, S., Van Lancker, V., Deleu, S., De Batist, M., and J.P. Henriet, 2003, Tertiary and Quaternary Geology of the Belgian Continental Shelf, SPSD II report, 77 pages.

Leeder, M.R. 1987. Tectonic and palaeogeographic model for Lower Carboniferous Europe; In: Miller, A.E. Adams and V.P. Wright (editors), European Dinantia Enviroment. Chichester, John Wiley & Sons, 1-10 p.

Lewis S., Maddy D., and Glenday S., 2004, The Thames valley sediment conveyor: fluvial system development over the last two interglacial-glacial cycles. Quaternaire, v. 15, p. 17–28.

Lunkka, J.P. 1994. Sedimentation and lithostratigraphy of the North Sea Drift and Lowerstoft Till Formation in the coastal cliffs of NE Norfolk, England, Journal of Quaternary Science, v. 9, 209–234.

Mansy, J.-L., Manby, G.M., Averbuch, O., Everaerts, M., Bergerat, F., Van Vliet-Lanoë, B., Lamarche, J. & Vandycke, S., 2003, Dynamics and inversion of the Mesozoic Basin of the Weald-Boulonnais area: role of basement reactivation, Tectonophysics, v. 373, p. 161–179.

Meijer, T. and Preece, R.C., 1995. Malacological evidence relating to the insularity of the British Isles during the Quaternary. Geological Society, London, Special Publications, p. 96, v. 89-110.

Murton, D.K. and Murton, J.B., 2012. Middle and Late Pleistocene glacial lakes of lowland Britain and the southern North Sea Basin. Quaternary International, v. 260, p. 115-142.

Patton, H., Hubbard, A., Andreassen, K., Auriac, A., Whitehouse, P. L., Stroeven, A. P., Shackleton.C., Winsborrow, M., Heyman, J., and Hall, A. M., 2017, Deglaciation of the Eurasian ice sheet complex, Quaternary Science Reviews, v. 169, p. 148–172.

Peeters, J., Busschers, F. S., and Stouthamer, E., 2015, Fluvial evolution of the Rhine during the last interglacial-glacial cycle in the southern North Sea basin: a review and look forward. Quaternary International, v. 357, p. 176–188.

Peeters, J., Busschers, F. S., Stouthamer, E., Bosch, J. H. A., Van den Berg, M. W., Wallinga, J., Versendaal, A.J., Bunnik, F.P.M., and Middelkoop, H., 2016, Sedimentary architecture and chronostratigraphy of a late Quaternary incised-valley fill: A case study of the late Middle and Late Pleistocene Rhine system in the Netherlands, Quaternary Science Reviews, v. 131, p. 211–236.

Reynaud, J-Y., Tessier, B., Auffret, J-P., Berné, S., De Batist, M., Marsset, T., and Walker, P., 2003, The offshore Quaternary sediment bodies of the English Channel and its Western Approaches, Journal of Quaternary Science, v. 18, p. 361–371.

Sejrup, H. P., Nygård, A., Hall, A. M., and Haflidason, H., 2009, Middle and Late Weichselian (Devensian) glaciation history of south-western Norway, North Sea and eastern UK, Quaternary Science Reviews, v. 28, p. 370-380.

Sejrup, H. P., Clark, C. D., and Hjelstuen, B. O., 2016, Rapid ice sheet retreat triggered by ice stream debuttressing: Evidence from the North Sea. Geology, v. 44, p. 355–358.

Smith, A.J., 1985, A catastrophic origin for the palaeovalley system of the eastern English Channel. Marine Geology, v. 64, p. 65–75.

Shennan, I., Lambeck, K., Flather, R., Horton, B., McArthur, J., Innes, J., Lloyd, J., Rutherford, M., and Wingfield, R., 2000. Modelling western North Sea palaeogeographies and tidal changes during the Holocene. In Shennan, I. & Andrews, J. (eds.), Holocene Land-Ocean Interaction and Environmental Change around the North Sea, Geological Society London Special Publication, v. 166, p. 299–319.

Stucchi, M., Rovida, A., Capera, A. G., Alexandre, P., Camelbeeck, T., Demircioglu, M. B., Gasperini, P, Kouskouna, V., Musson, R.M.W., Radulian, M., Sesetyan, K., Vilanova, S., Baumont, D., Bungum, H., Fäh, D., Lenhardt, W., Makropoulos, K., Martinez Solarez, J.M., Scotti, O., Živčić, M., Albini, P., Batllo, J., Papaioannou, C., Tatevossian, R., Locati, M., Meletti, C., Vigano, D., and Giardini, D., 2013, The SHARE European earthquake catalogue (SHEEC) 1000–1899, Journal of Seismology, v. 17, p. 523–544.

Sturt, F., Garrow, D. and Bradley, S., 2013, New models of North West European Holocene palaeogeography and inundation, Journal of Archaeological Science, v. 40, p. 3963–3976.

Tappin, D.R., 1991, Spurn Quaternary geology, Sheet 53°N–00°, British Geological Survey, scale 1:250 000, 1 sheet.

Ter Wee, W.X, 1983, The Elsterian glaciation in the Netherlands. In J. Ehlers (ed.), Glacial Deposits in North West Europe, p. 413–415. Rotterdam: Balkema

Toucanne, S., Zaragosi, S., Bourillet, J. F., Gibbard, P. L., Eynaud, F., Giraudeau, J., Turon, J.L., Cremer, M., Cortijo, E., Martinez, P., and Rossignol, L., 2009a, A 1.2 Ma record of glaciation and fluvial discharge from the West European Atlantic margin. Quaternary Science Reviews, v. 28, p. 2974–2981.

Toucanne, S., Zaragosi, S., Bourillet, J.F., Cremer, M., Eynaud, F., Van Vliet-Lanoë, B., Penaud, A., Fontanier, C., Turon, J.L., Cortijo, E. and Gibbard, P.L., 2009b, Timing of massive 'Fleuve Manche' discharges over the last 350 kyr: insights into the European ice-sheet oscillations and the European drainage network from MIS 10 to 2, Quaternary Science Reviews, v. 28, p. 1238–1256.

Toucanne, S., Zaragosi, S., Bourillet, J. F., Marieu, V., Cremer, M., Kageyama, M., Van Vliet-Lanoë, B., Eynaud, F., Turon, J-L, and Gibbard, P. L., 2010, The first estimation of Fleuve Manche palaeoriver discharge during the last deglaciation: evidence for Fennoscandian ice sheet meltwater flow in the English Channel ca 20–18ka ago, Earth and Planetary Science Letters, v. 290, v. 459–473.

Toucanne, S., Soulet, G., Freslon, N., Jacinto, R. S., Dennielou, B., Zaragosi, S., Eynaud, F., Bourillet, J-F., and Bayon, G., 2015, Millennial-scale fluctuations of the European Ice Sheet at the end of the last glacial, and their potential impact on global climate. Quaternary Science Reviews, v. 123, p. 113–133.

Turner, C., 2000, The Eemian interglacial in the North European plain and adjacent areas. Netherlands Journal of Geosciences, v. 79, p. 217–231

Van Vliet-Lanoë, B. et al., 2002, Cenozoic evolution of the Channel and southern North Sea areas (Western Europe). The neotectonic control on sedimentation and rivers; in: Contributions to the Geology of Belgium and Northwest Europe: Proceedings of the First Geologica Belgica International Meeting, Leuven, v. 12, p. 21–24, editors: Degryse, P. & Sintubin M.

Zalasiewicz, J.A., and Balson, P.S., 1985, East Anglia sheet 52°N–00°, Solid Geology, British Geological Survey, 1:250 000 series, 1 sheet.

Zagwijn, W.H., 1989, The Netherlands during the Tertiary and the Quaternary: a case history of Coastal Lowland evolution. In Coastal Lowlands, p. 107–120. Springer Netherlands.

Ziegler, P.A., 1990, Geological Atlas of Western and Central Europe (2nd edition), Amsterdam. 778 Shell Internationale Petroleum Maatschappij BV.

Chapter 3

Methods

Chapter 3 – Methods

The present study is based on the analysis and interpretation of marine geophysical data. These consist of bathymetric data and 2-dimensional single-channel and multi-channel seismic-reflection profiles (see Table 3.1 and Table 3.2). The geophysical dataset gathered for this study comprises seismic reflection data acquired by the author of this study and colleagues during geophysical surveys undertaken in 2010, 2012 and 2016; bathymetric and seismic reflection datasets provided by various British, Dutch and French organizations, involving data acquired since the early 1980's to 2013; and data collected between 1980 and 2013 stored at the Renard Center of Marine Geology (RCMG – Faculty of Science, Department of Geology, Ghent University).

The interpretation of the geophysical data was carried out by combining the 2D and 3D seismic interpretation software Opendtect and IHS Kingdom Suite, with Geographic Information System (GIS) ArcMap and Global Mapper. Seismic-stratigraphy, velocity models and geological mapping have been aided by combining the seismic reflection and bathymetric data with a series of sediment-core descriptions furnished by the archives of the French Bureau de Recherches Géologiques et Minières (BRGM; www.brgm.fr) and SediLITHO@SEA (cf. EU FP7 Geo-Seas; www.geoseas.eu databases). See tables in appendix A attached to this manuscript and maps in Chapter 4 and in De Clercq et al. (2016).

3.1 Seismic reflection data

Seismic reflection data consist of single-channel and multi-channel 2D seismic-reflection profiles collected from the Dover Strait, and single-channel high-resolution seismic-reflection profiles acquired from the Belgian Continental Shelf (BCS) and Outer Thames Estuary (OTE). We have divided these data in two datasets; i.e. the Dover Strait dataset, used mostly for the characterization of the geology and geomorphology of the Dover Strait (Chapters 4, 5 and 6), and the southern North Sea dataset, employed for the geomorphological analysis of the southern North Sea (Chapters 7).

The majority of the seismic reflection datasets gathered for the present study consist of data acquired by means of Sparker, GI-Guns and Boomer sources in combination with either single-channel or multi-channel streamers (Figure 3.1 and Figure 3.2).

Name	Year	Acquired by	Processed at	V-Resolution	Max.Penetration
Dangeard I	2002	Lille U. – RCMG	ROB	1–3 m	80–100 m
ROB–RCMG	2010 & 2012	ROB–RCMG	ROB	Sc: 1–3 m; Mc: 5–10 m	Sc: 100–150 m Mc: 250 m
BCS	1980–2016	Various sources	RCMG	0.15–3 m	15–100 m
OTE REC	2007	Emu Ltd–IECS-UH	Emu Ltd–IECS-UH	< 1 m	70–80 m

Table 3.1. Seismic reflection datasets available for the present study. Lille U.: Lille 1 University; RCMG: Renard Center of Marine Geology; ROB: Royal Observatory of Belgium; Emu Ltd–IECS-UH: Emu Ltd and the Institute of Estuarine Coastal Studies (University of Hull). OTE: Outer Thames Estuary; BCS: Belgian Continental Shelf; Sc: single-channel seismic-reflection data; Mc: Multi-channel seismic-reflection data.

Sparker sources produce acoustic energy by electric discharging through multiple electrodes. The energy used ranges from 100 to 600 J (see Reynolds, 2011). The electric discharge ionizes the seawater between the tips of the electrodes and the frame, creating hydrogen bubbles. These bubbles implode generating high-frequency acoustic waves (Reynolds, 2011). The collapsing bubbles produce a broad band (50 Hz – 4000 Hz) omnidirectional pulse, which can penetrate several hundred meters into the subsurface. Seismic reflection data are the result of multiple reflections of the acoustic waves as they traverse geological units with different acoustic impedances (Reynolds, 2011). The higher the difference in acoustic impedance is, the higher the reflection coefficient will be, and the more distinct the reflections will be in the seismic reflection profile (Reynolds, 2011). The vertical scale in seismic reflection profiles is determined by the time span between the emission and reception of the signal, and is typically expressed in Two-Way Time (TWT). Penetration depths depend on the energy and frequency content of the signal emitted, as well as on the distance between sources and receivers. Time–depth conversions are performed by applying, if known, the various sound velocities that characterize each of the units composing the substratum. Vertical resolution depends on the wavelength of the acoustic signal, energy scattering and spherical divergence. Horizontal resolution depends on the cruise speed and shot intervals, as well as the Fresnel Zone (see Reynolds, 2011).

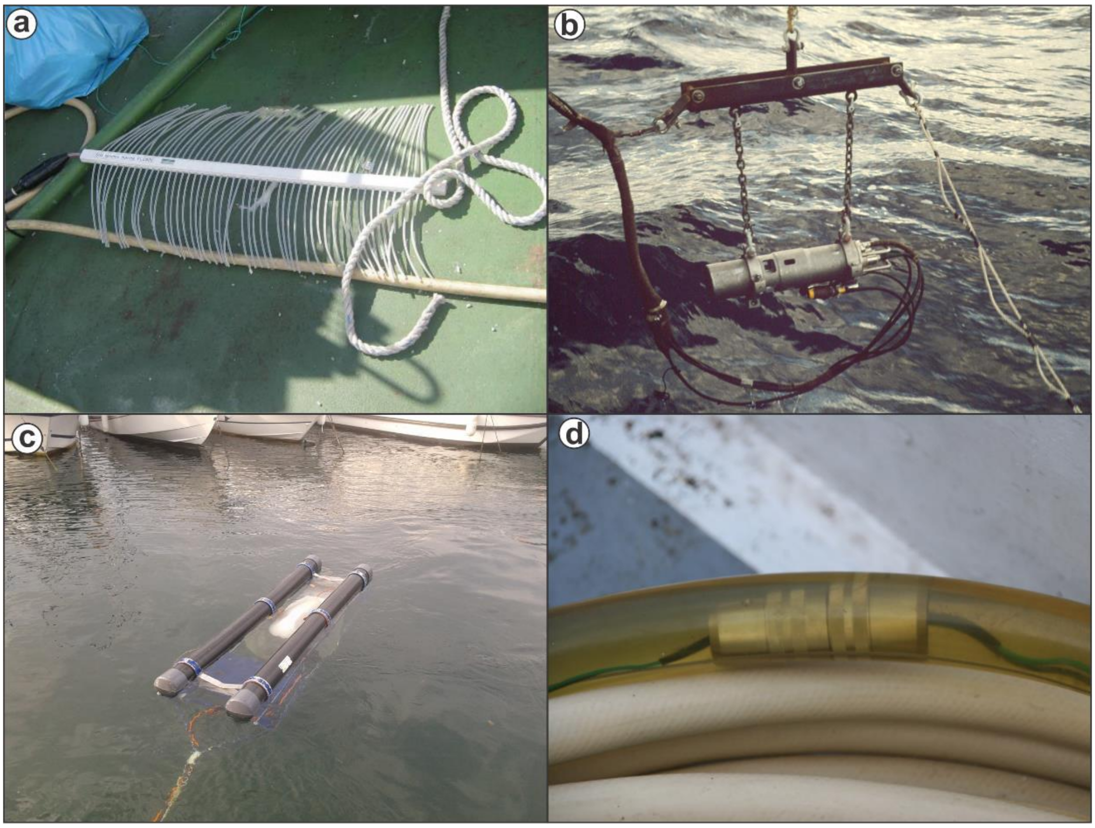

Figure 3.1. Seismic sources. a) Centipede Sparker; b) GI-Gun; c) Boomer; d) hydrophone within a single-channel streamer.

Boomer sources also store electrical energy in capacitors, but they discharge through a flat spiral coil instead of generating a spark. By sending electrical energy from the power supply through the wire coils, the two spring-loaded plates in the boomer transducer are electrically charged causing the plates to repel, thus generating an acoustic pulse. The boomer is a broad-band acoustic source operating in the 300 Hz – 3000 Hz range. Boomer resolution usually ranges from 0.5 to 1 m and penetration depth ranges from 25 up to 80 m.

GI-gun sources consist of one or more pneumatic chambers that are pressurized with compressed air surrounding a piston/shuttle. Acoustic waves are generated by a release of a specified volume of air into the water at regular time intervals. The air release coalesces into a bubble, thereby generating sound by the ensuing expansions and contractions of the bubble. GI gun arrays are typically designed to create source wavelets that are as compact in time as possible and that have minimal bubble oscillations. Compact wavelets have indeed wide signal-frequency spectra, and the less bubble oscillations there are, the smoother the signal spectrum will be. The air guns are relatively deep penetration sources, operating at 100 to about 1200 Hz.

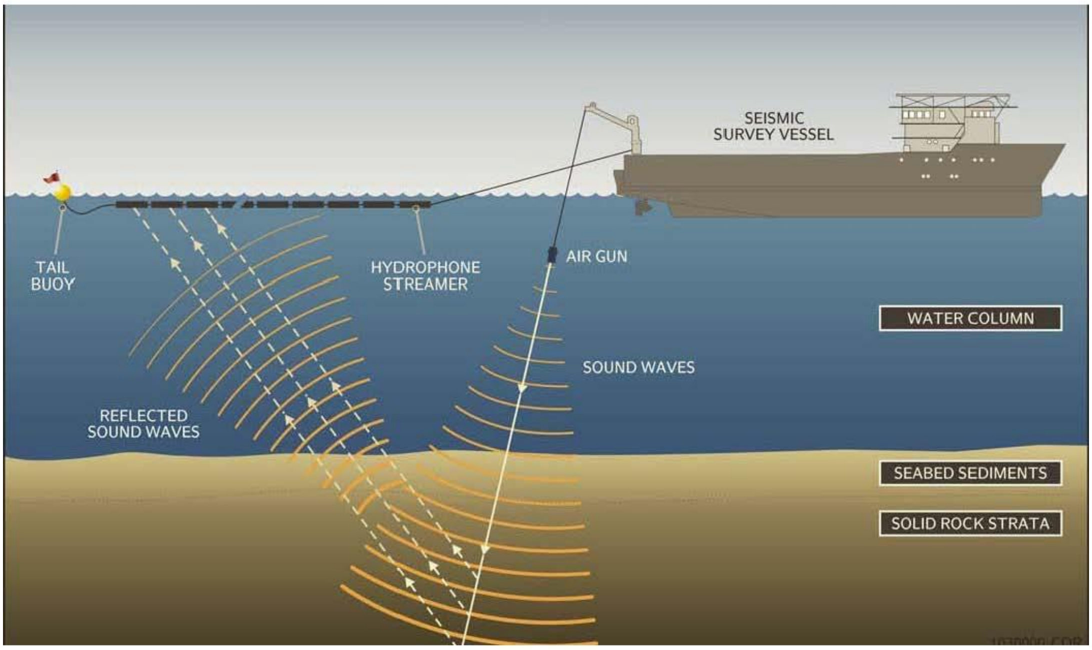

Figure 3.2. Cartoon showing a typical array during offshore seismic reflection surveys.

3.1.1. Dover Strait seismic reflection dataset

The seismic reflection data available from the Dover Strait consist of ~515 line-kilometer of 2D single-channel seismic-reflection profiles and ~200 line-kilometer of 2D deeper-penetration multi-channel seismic-reflection profiles (Figure 3.3). These data were acquired along the entire width of the Dover Strait during 4 geophysical campaigns; i.e. the Dangeard 1 campaign, and the 2010/09, 2012/03 and 2012/25 RV Belgica Campaigns (Table 3.1). Acquisition reports are available at https://odnature.naturalsciences.be/belgica/en/index.

3.1.1.1. Acquisition

The Dangeard 1 campaign was carried out in 2002 on board of the research vessel INSU "Côtes de la Manche" by a collaboration of RCMG and Lille 1 University. During this survey, data acquisition was performed by using a 300 J Centipede Sparker system triggered every 1 s, and a single-channel surface streamer consisting of 10 hydrophones. Both seismic and positioning data were logged with Delph iXBlue software. This dataset consists of 20 parallel, ~10 km long profiles acquired with ~200 m spacing, and one cross-profile. Those data exhibit vertical and horizontal resolutions better than 1–3 m. Maximum penetration into substratum ranges between 80–100 m before the seismic multiples obscure the signal (Table 3.1).

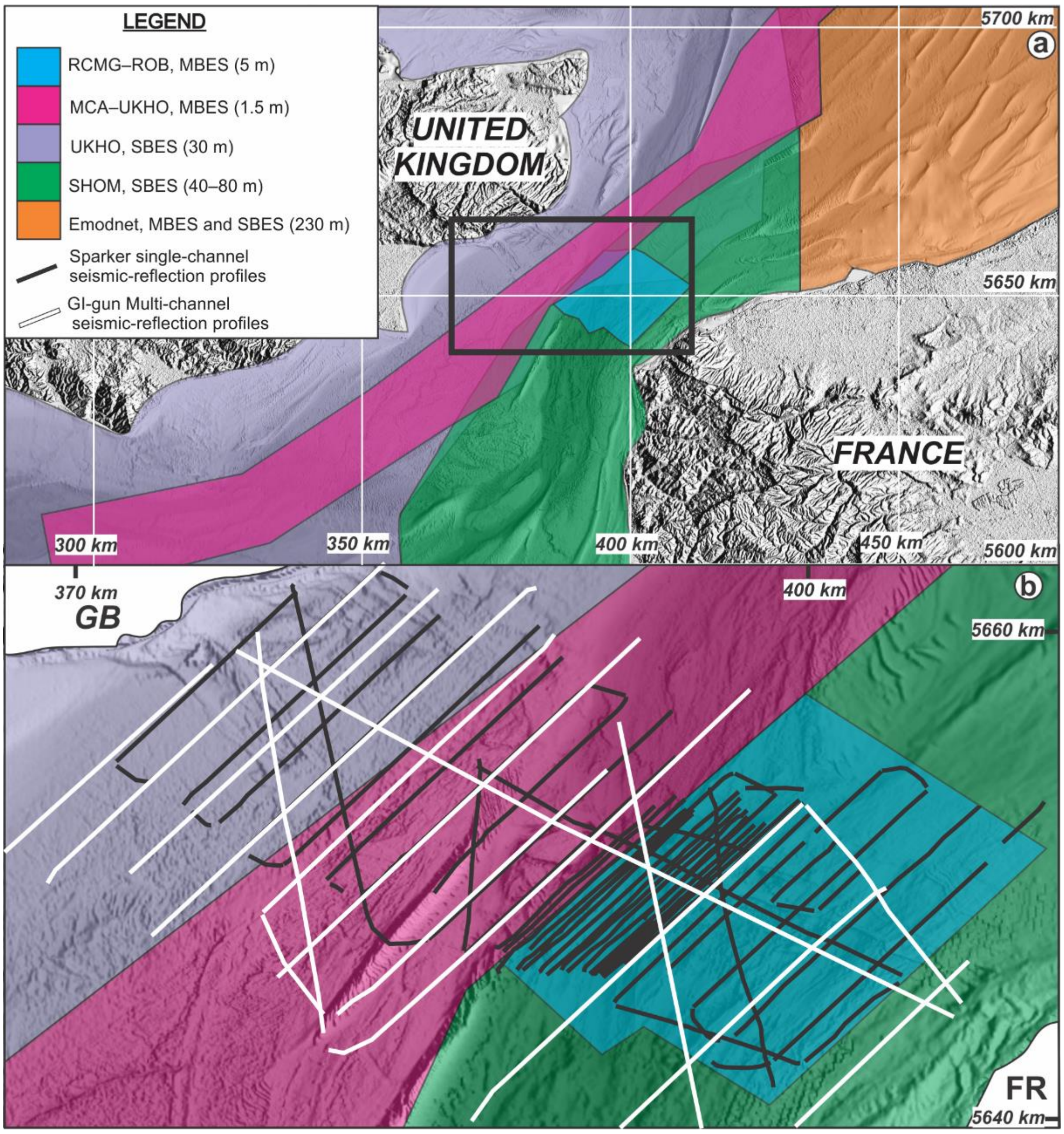

Figure 3.3. Dover Strait bathymetric and seismic reflection datasets. Projections: UTM zone 31N/WGS84; distances are given in kilometers.

The 2010/09 and 2012/03 RV Belgica cruises were performed by the Royal Observatory of Belgium (ROB) and RCMG on board of the research vessel Belgica. Data acquisition was conducted by means of a 500 J Centipede sparker system triggered every 1 s, and a single-channel surface streamer of 10 hydrophones. Data were recorded through a Delph IXSEA seismic acquisition software system. Sampling frequency was set at 10 kHz and record length at 0.4 s TWT. Together, the two surveys consist of 43 seismic-reflection profiles spaced 1.5 km apart and 5 cross-profiles. These surveys were performed to the southeast and northwest of the 2002 seismic survey (Figure 3.3). This dataset exhibits vertical and horizontal resolutions better than 1–3 m. Maximum penetration into the substratum ranges between 80–150 m before the seismic multiples obscure the signal (Table 3.1).

The 2012/25 Belgica Campaign was conducted by the ROB and RCMG with technical assistance and equipment provided by the Royal Netherlands Institute for Sea Research (NIOZ). During this campaign, data were acquired by means of a GI-gun as a source (Figure 3.1b), and a receiver consisting of a 24-channel streamer composed of 10 hydrophones per channel (channel spacing: 10.5 m). GI-gun was fired at ∼100 bars of pressure every 7 s. The sampling frequency was set to 1000 Hz, for the first 9 lines, and 2000 Hz, for the rest; during those two phases, the record length was set to 5 and 2 s, respectively. In total, we acquired 12 profiles sub-parallel to the English and French coasts and 4 control cross profiles. These profiles have lengths ranging between 13 km and 20 km. During acquisition, parallel profiles were spaced 2–5 km to one another (Figure 3.3). These data penetrate up to ∼250 m into the substratum before the multiples obscure the signal, and exhibit vertical resolution ranging from 5 m to 10 m.

3.1.1.2. Processing

Single-channel seismic-reflection profiles were processed at the ROB (Figure 3.4) by Dr. K. Vanneste with contributions from the author of this dissertation for tide corrections. Processing was performed by means of the open-source package Seismic Unix (SU) (Stockwell 1999) in conjunction with a python interface to SU (Supy) developed at the ROB. Processing consisted of (see Vanneste et al., 2011):

1) Swell Filter. When the movements of the source and receiver are larger than the wavelength of the signal produced by the source, seismic reflectors do not line up. To improve reflection coherency, a swell filter must be applied. Swell filtering consists of detecting the seafloor in each trace, smoothing (moving average) the detected seafloor, calculating the time differences between the original and the smoothed seafloor, and applying these shifts.

2) Navigation Correction. Positioning during navigation was recorded in geographic coordinates. These were converted to a metric coordinate system (UTM, Zone 31N), and written in the trace headers.

3) Predictive Error Filtering. The reflected signal is the result of convolution of the source wavelet with the substratum’s reflection series. Theoretically, it should be possible to retrieve the substratum’s reflection series by applying spiking

deconvolution, collapsing the source signal into a spike. One way to achieve this is by applying a Predictive Error (Wiener) Filtering with zero gap and a lag (prediction distance), corresponding to the length of the source signal.

4) Band-pass Filtering and Time-varying Gain. This filter is applied to remove low-frequency and high-frequency noise, and to recover the amplitude of deeper reflections. We bandpass-filtered the data between 250 and 750 Hz and applied time-varying gain.

5) Muting. In this step, the amplitude of the signal before the seafloor arrival is set to zero. That eliminates the direct arrival (signal travelling directly from source to receiver) and the various diffractions produced in the water column.

6) Depth conversion and depth migration. In this step, time sections were converted into depth profiles. Simple depth conversion consists of resampling from equidistant time intervals to equidistant depth intervals. Simple vertical 'stretching' of seismic times to seismic depths cannot correct for lateral position errors that may be present in the seismic image. To address this problem, the section must be migrated. Migration is a seismic processing step to reposition reflections at their correct surface location. Sediment cores were only available in the chalk and mudstone Cretaceous sedimentary formations. However, seismic reflection profiles traverse both laterally and vertically several other sedimentary units of different composition and compaction. This precluded accurate lateral and vertical velocity analyses. Depth conversions were then performed by assuming a uniform sound velocity of 2000 ms^{-1} below the seafloor, and 1500 ms^{-1} through the water column. This approach appears to be acceptable in most cases, but may not be appropriate in places where channel-fill deposits occur below the seafloor.

7) Tide correction. Tide corrections were performed by applying the tide information provided by the British National Oceanography Centre and the Hydrographic and Oceanographic Service of the French Marine (SHOM). The exact times of the shots during data acquisition were not available for the Dangeard I Cruise. Therefore, it was impossible to know the exact height of the tides at the different shots' location. Tide corrections were in that case applied by extracting the seafloor height of every

seismic trace from the 1.5 m bin-size tide-corrected bathymetric data available for this area (see section 3.2).

8) Data archive. Data was exported in SEG-Y format so it could be later read on most interpretation softwares.

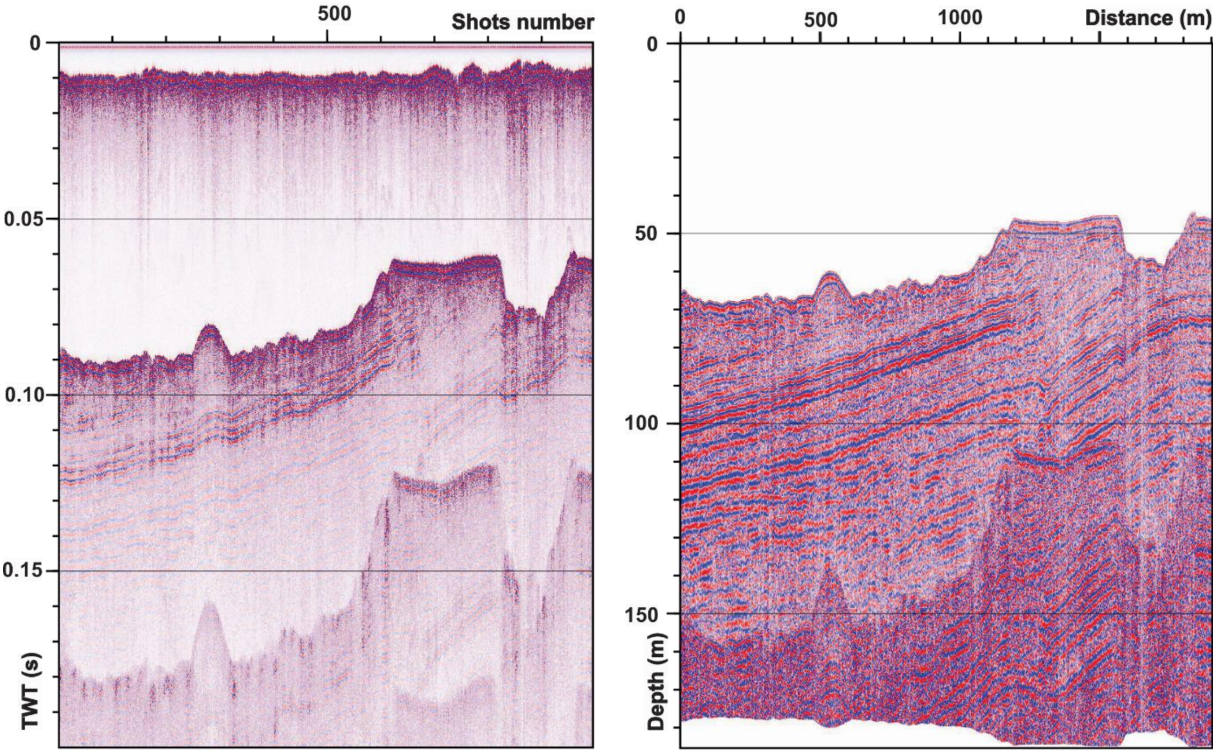

Figure 3.4. Left, Single-channel seismic-reflection profile before processing. Right, same seismic reflection profile after processing.

Multi-channel seismic-reflection profiles were processed at the RCMG by O. Zurita-Hurtado with contributions from the author of this dissertation. Processing was carried out by means of the processing software RadExPro (www.radexpro.com). Processing consisted of (see Figure 3.5):

1) 2-Dimensional-geometry Correction. Positioning during navigation was recorded in geographic coordinates. These were converted to a metric coordinate system (UTM, Zone 31), and written in the trace headers.

2) Trace Editing. Noisy and/or dead traces were edited or deleted.

3) Swell Filtering. A swell filter was applied to attenuate heave effects. This helps to flatten the seabed and the subsequent horizons.

4) Spherical Divergence Correction. Gain was applied to reduce the attenuation of the

amplitudes of the reflections with depth (later arrivals).

5) Band pass filter. Removal of noisy frequencies (18, 30, 325, 375 Hz).

6) FK filter. Removal of side scattered noise and spatially random noise in the shot domain.

7) 2:1 channel interpolation.

8) Pre-stack multiple removal in the shot domain. Multiples are delayed reflections that interfere with the "primary" reflections we want to image. This delay occurs due to the recording of reflection energy that has taken a more complex and longer ray path from source to receiver than primary reflections.

9) Velocity analysis every 50 Common Mid Points (CMP). This step is intended to create a model of the seismic velocities through the sub-surface. It is performed in the Common Depth Point (CDP) domain, where the assumption of hyperbolic move-out of reflections is often reasonable. The aim of the velocity analysis is to find the velocity that flattens a reflection hyperbola and thus provides the best result when stacking is applied (Yilmaz, 2001).

10) Normal move-out (NMO) correction. The NMO correction is applied to correct hyperbolic-curve artefacts produced by delay in travel time.

11) Stack. All the traces corresponding to the same CMP are summed up into one trace, thus generating a pseudo zero-offset section (similar to the one obtained from a single-channel streamer).

12) Post-stack multiple removal.

13) Post-stack random noise suppression. Removal of residual side scattered noise and spatially random noise.

14) Post-stack amplitude scaling and lateral amplitude compensation.

15) Muting. Data above water bottom was removed.

16) Data archive. Data was exported in SEG-Y format so it could be later read on most seismic interpretation software.

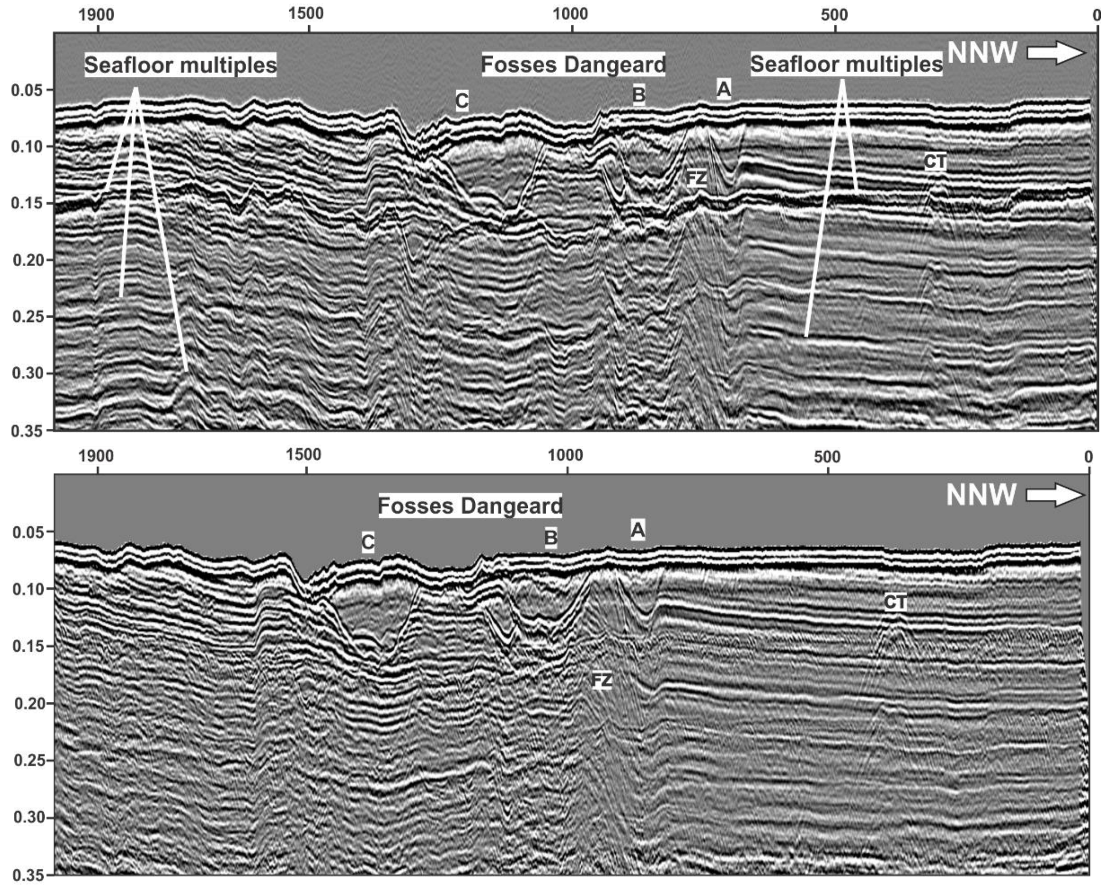

Figure 3.5. Multi-channel seismic-reflection profile acquired across the Fosses Dangeard. Above, stack generated prior to processing step "Pre-Stack multiple removal". Below, final processed output. Fosses Dangeard (A, B and C) traversed by this profiles are indicated. CT: hyperbolic diffraction due to the Channel Tunnel; FZ: Reverse fault zone. For more information on the labelled features see next Chapters.

We were not able to record the far-field signature during data acquisition and, hence, the source wavelet was not correctly extracted from this data. Moreover, the combination of shallow waters and hard seabed produced strong multiple reflections that are very difficult to eliminate. Different statistical deconvolution techniques were tested, but they proved ineffective. Consequently, we were able to attenuate the first multiple, but not to eliminate it (Figure 3.5). Further, primary reflections deeper than ~0.25 s were completely obscured by high-order multiples, preventing the correct interpretation of the deep geology and structure. The quality of the processed data was however high enough to complete the 3-dimensional characterization of the Fosses Dangeard (see Figure 3.5). These data were also used to aid the characterization of the geometry of faults down to 200 m deep and their lateral extents.

3.1.2. Southern North Sea seismic reflection dataset

Southern North Sea Seismic reflection data comprise several seismic reflection datasets acquired from the Belgian Continental Shelf (BCS) and Outer Thames Estuary (OTE) areas (Figure 3.6).

3.1.2.1. BCS seismic reflection dataset

The BCS dataset covers the entire extent of the Belgian Continental Shelf (i.e. 3500 km^2). It is mainly composed of the dataset gathered by De Clercq et al. (2016), which included the 5300 line-kilometers of 2D seismic-reflection data gathered by Mathys (2009a), and 1655 line-kilometers of 2D seismic-reflection data acquired in 2008 and between 2013 and 2015 (Figure 3.6). The De Clercq et al. (2016) dataset was completed with 665 line-kilometers of seismic reflection profiles collected from the northwestern part of the BCS by the author of this manuscript and colleagues during two geophysical campaigns undertaken in 2016 (Figure 3.6).

The data gathered by Mathys (2009a) consist of several seismic reflection datasets collected between 1980 and 2007. These data were acquired mainly using Sparker and Boomer sources. Vertical resolutions of this dataset go from 0.5 m to 3 m and penetrations range between 12 m to 80 m. Four thousand line-kilometers of this database were digitized from paper seismic-reflection profiles and converted to SEGY. Technical specifications on the acquisition, processing and digitization of these data were extensively described in Mathys (2009a)'s PhD dissertation.

De Clercq et al. (2016) reanalyzed Mathys (2009a) database and complemented it with 750 line-kilometers of seismic profiles collected from the Hinderbank region in 2008 and 905 line-kilometers acquired during 7 geophysical surveys performed between 2013 and 2015. The Hinderbank seismic reflection profiles were acquired with a Boomer source and a single-channel streamer. These data have resolutions better than 0.5–1 m and penetrations of ~30 m. Technical specifications on this dataset acquisition and processing were described in reports Depret-G-tec (2009) and Mathys et al. (2009b).

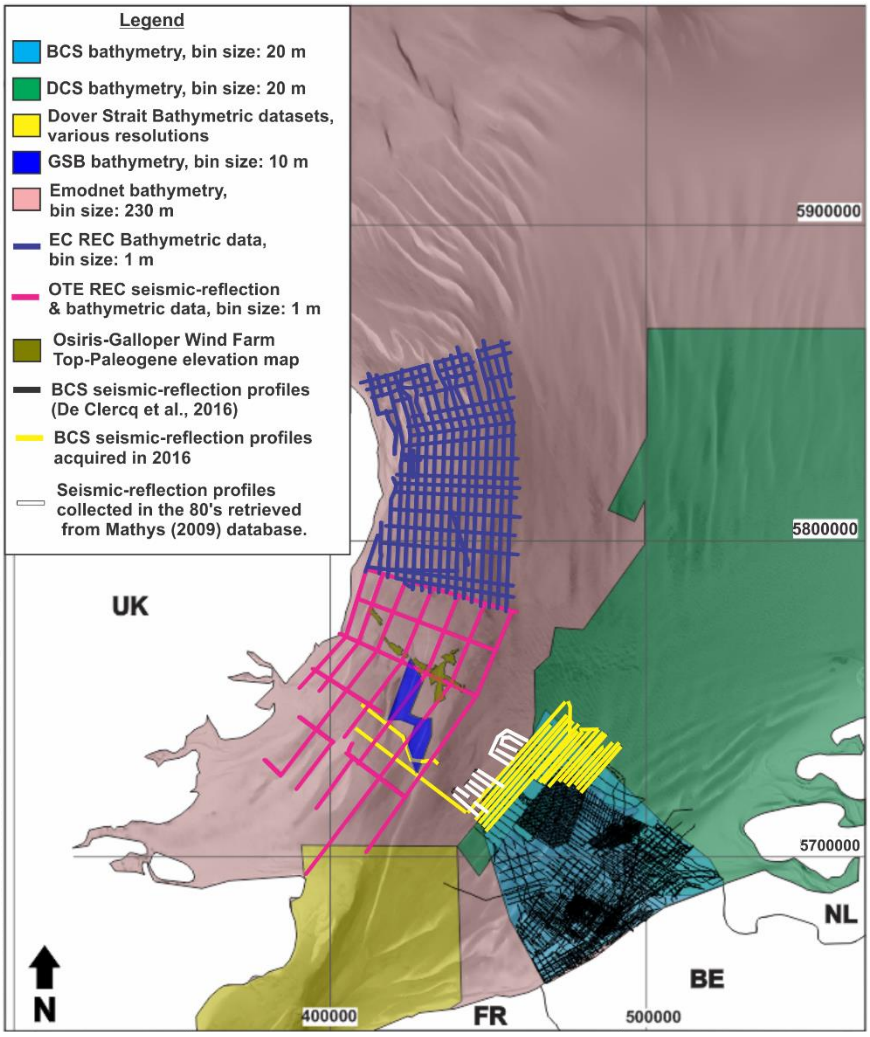

Figure 3.6. Southern North Sea bathymetric and seismic reflection datasets.

Seismic data collected between 2013 and 2015, as well as those of 2016, were acquired on board of the RV Simon Stevin managed by the Vlaams Instituut voor de Zee (VLIZ). These surveys were undertaken within the framework of the RCMG–Deltares–VLIZ collaborative project "IWT SBO–120003 'SeArch': Archaeological heritage in the North Sea". This dataset includes data acquired with various sources and both single-channel and multichannel receivers. Vertical resolution ranges from 0.15 m to 1 m, and penetrations from 15 m to a few hundred meters. Technical specifications on the acquisition and processing are described in report "Synthese Sub-seafloor imaging – WP1.3.4", available at http://www.sea-arch.be/nl/resultaten. Reports on the data acquisition during surveys can also be consulted at http://www.sea-arch.be/nl/resultaten (reports titles: WP1.2.3_A to WP1.2.3_H).

In the present study, we do not interpret the geologic structure and seismic-stratigraphy imaged on the BCS seismic reflection profiles. Rather, we focus on the geomorphology of the surface defining the basal erosional surface of Quaternary sediments, which in the BCS correspond to the top-Paleogene angular unconformity. The acoustic signature of that surface is very distinct in the seismic reflection data, as it forms a prominent angular unconformity and usually exhibits high-amplitude reflectivity. This surface was picked and correlated among seismic reflection profiles by M. De Clercq and the author of this dissertation, and depth converted by V. Chademenos. The depths along this surface were modeled in meters by calculating average velocities between the seafloor and the top pre-Quaternary surface. This was done by comparing the vertical distance between these two horizons in the seismic reflection data with the vertical positions of these surfaces in a series of boreholes (cf. EU FP7 Geo-Seas; www.geoseas.eu) drilled along some of the seismic reflection lines (see boreholes' locations in Clercq et al., 2016). The depth-converted 2-dimensional seismic horizon defining the base-Quaternary surface was picked in the various seismic profiles and gridded in order to produce a depth-converted structure map modeling the elevation of that surface relative to the sea level (Lowest Astronomic Tide – LAT). In this dissertation we refer to that elevation map as top pre-Quaternary surface. Note that in this study, we have extended the top pre-Quaternary structure map produced by De Clercq et al. (2016) further north by adding interpretations from the data that we collected in 2016. We have also included in the new depth-converted structure map, a series of seismic reflection profiles from Mathys (2009a)'s database, which were not used in previous models (white lines in Figure 3.6).

The top pre-Quaternary surface derived from the combination of the various datasets available from the BCS was gridded at 25 m pixel size in order to show areas with no data. Boreholes were not available for the data acquired outside the BCS. We therefore applied a constant sound velocity for the seismic unit(s) between the seafloor and the top pre-Quaternary surface of 1560 ms^{-1}. This velocity represents the average velocity obtained from areas with available sedimentary information. The lack of boreholes to calibrate the latter time–depth conversion may have added some distortion to the depth-converted structure map in those areas. Based on several tests performed in areas with borehole information, we estimate that this distortion is limited to 0–1 m in areas where the thickness of this sediment package does not exceed ~10 m, which is the case for most of the area where we apply constant velocity. However, in localized areas covered by thick sandbanks, the sediment package between these surfaces can reach up to 20–30 m in thickness, which may induce local distortions of 1–2 m. This distortion is still acceptable for our purposes, as the geomorphological features for which we use those data are several orders of magnitude larger than the errors induced by this technique.

3.1.2.2. OTE seismic reflection dataset

The OTE seismic reflection dataset consists of ~700 line-kilometers high-resolution seismic-reflection profiles acquired with a Boomer source with a line spacing of 10–20 km (Figure 3.3). This dataset also includes localized depth-converted structure maps modeling the elevation with respect to the LAT of the basal erosional surface of Middle–Late Pleistocene sediments, which, in that area, are directly on top of Neogene and Paleogene substratum (Figure 3.6). In this study, we refer to that surface as top pre-Quaternary erosional surface too.

The seismic reflection profiles were acquired in 2007 by Emu Ltd and the Institute of Estuarine Coastal Studies of University of Hull within the Outer Thames Estuary Regional Environmental Characterization (data and reports available at marinedataexchange.co.uk). These data have maximum penetration before the seismic multiples obscure the signal of 70–80 m, and resolutions of less than 1 m. The data were provided as raw data. Apart from applying a band-pass filter (250–4000 Hz) and an automatic gain control, this dataset has not been processed.

The local depth-converted structure maps modelling the top pre-Quaternary surface were downloaded from marinedataexchange.co.uk as 11.7 m cell-size Surfer grids. These surfaces were derived from fully processed and tide-corrected seismic-reflection data collected in 2009 within the Osiris–Galloper Wind Farm site characterization project (see www.galloperwindfarm.com). Data and reports are available at marinedataexchange.co.uk.

3.2 Bathymetric data

The bathymetric data consist of a series of datasets with different resolutions (Table 3.2). These data can be divided in two main datasets; i.e. the Dover Strait dataset (Figure 3.3) and the southern North Sea dataset (Figure 3.6), both including single-beam and multibeam bathymetry.

dataset	Year	Type of data	Processed at/by	Gridded at
ROB–RCMG	2010–2012	MBES	RCMG	5 m
MCA–UKHO	2006–2007	MBES	MCA	1.5 m
UKHO	1988–2004	SBES	MCA – IC	30 m
SHOM	Since 1970's	SBES	Lille U.	40 m and 80 m
Emodnet	1946–2017	SBES & MBES	Multiple Inst.	230 m
BCS	1990–2015	SBES & MBES	FHS	20 m
DCS	?–2013	SBES & MBES	Multiple Inst.	25 m
GSB	2005	MBES	GGOW	10 m
OTE REC	2008	SSS & HRSRP	Emu Ltd–IECS-UH	1 m
EC REC	2008	SSS & HRSRP	Limpenny et al. (2011)	1m

Table 3.2. Bathymetric datasets available for the present study. BCS: Belgian Continental Shelf; DCS: Dutch Continental Shelf; GSB: Gabbard Sandbank Bathymetry; OTE REC: Outer Thames Estuary Regional Environmental Characterization; EC REC: East Coast Regional Environmental Characterization; MBES: Multibeam echosoundings; SBES: Single-beam echosoundings; SSS: side-scan sonar; HRSRP: High-resolution seismic-reflection profiles; Lille U.: Lille 1 University; RCMG: Renard Center of Marine Geology; ROB: Royal Observatory of Belgium; FHS: Flemish Hydrographic Service; MCA: Maritime and Coastguard Agency; Multiple Inst.: Multiple institutions; GGOW: Greater Gabbard Offshore Wind; IC: Imperial College London. See Table 3.1 for other labels.

Single-beam echosounder systems are generally configured with a transceiver (transducer/receiver). The transceiver contains a transmitter, which controls pulse length and provides electrical power at a given frequency. It transmits a high-frequency acoustic pulse in a beam directly downward into the water column (Figure 3.7b). Acoustic energy is reflected off the seafloor beneath the vessel and received again at the transceiver. This transmit-receive cycle repeats at a fast rate, on the order of 1 kHz. The continuous recording of the time difference between the transmission and the reception of the acoustic signal yields high-resolution topographic profiles along the survey tracks. Original data thus provide depths in two-way time (TWT), which is converted to depth in meters by applying the sound velocity in water. Sound velocity in water is a function of the temperature, salinity and pressure/depth ratio of the water. These parameters can be measured by performing sound velocity profiles before or during the data acquisition.

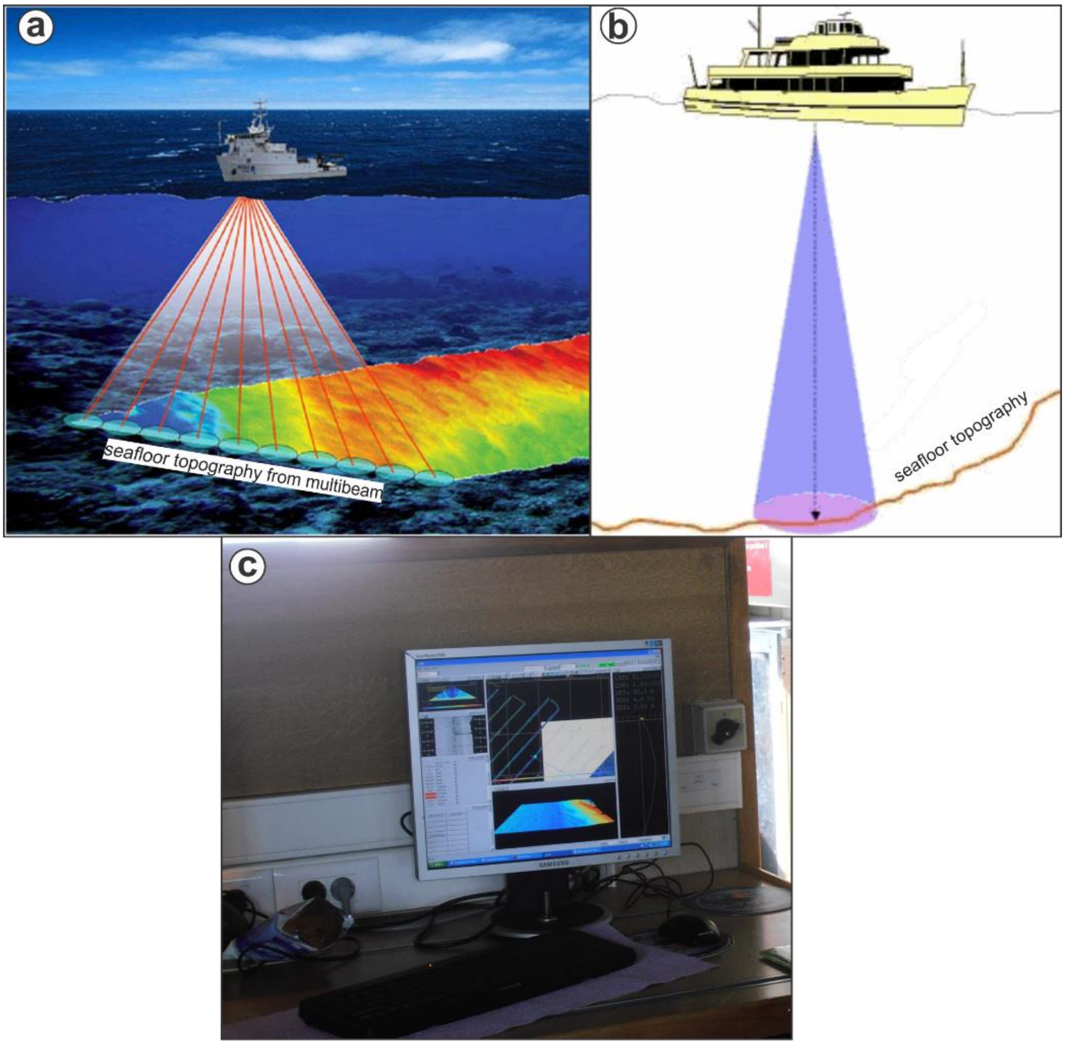

Figure 3.7. a) Cartoon showing multibeam swath bathymetry acquisition; b) cartoon showing single-beam bathymetry acquisition; c) picture showing monitoring during data acquisition with Kongsberg EM3002 multibeam echosounder.

Multibeam echosounders emit sound waves in the shape of a fan from directly beneath a ship's hull. These systems measure and record the time it takes for the acoustic signal to travel from the transmitter (transducer) to the seafloor and back to the receiver. In this way, multibeam sonars produce a “swath” of soundings (i.e. depths) for broad coverage of a survey area (Figure 3.7a; e.g. Hughes-Clarke et al., 1996). The coverage area depends on the depth of the water, and typically ranges between two and four times the water depth. Various transmitted frequencies are utilized by different MBES systems depending on the seafloor depth. Many multibeam systems are capable of recording additional acoustic backscatter data. Multibeam backscatter is intensity data that can be processed to create a low-resolution image of the seafloor. Backscatter is co-recorded with bathymetric data and is often used to assist with bathymetric data interpretation and post-processing.

3.2.1 Dover Strait bathymetric dataset

The Dover Strait dataset includes 4 bathymetric sets acquired and processed at different times by several organizations (Table 3.2). These data were gathered in collaboration with our colleagues from Imperial College London (see Chapters 4 to 6). The various datasets were referenced to the ITRF 2000 datum, used by the MCA-UKHO MBES dataset. Each dataset was also converted to UTM, Zone 31N coordinates. Areas that were not covered by these bathymetric surveys have been filled up with freely available online data downloaded from the European Marine Observation and Data Network (www.emodnet-hydrography.eu), which provides bathymetric data gridded at 230 m cell size.

3.2.1.1 ROB–RCMG MBES dataset

The ROB–RCMG multibeam bathymetric echosoundings (MBES) were acquired specifically for this PhD study by the author of this manuscript and colleagues within the framework of a collaborative project between the RCMG and ROB. These data consist of two datasets acquired during RV Belgica cruises 2010/09 and 2012/03 (see section 3.1.1 of this Chapter). They were acquired using a Kongsberg EM3002 multibeam echosounder (operating frequency of 300 kHz) installed permanently on the RV Belgica by the Belgian Ministry of Economic Affairs. Based on the sound absorption coefficient estimated during these campaigns and a sound profile that was taken with the SV-plus sound velocity probe on 15.04.2010 at 51.025°N, 1.51°E, an average sound velocity of 1,483 ms^{-1} was selected for the water layer. During surveys, data were collected keeping 10–20% of overlap between consecutive swaths. Acquisition reports are available at www.odnature.natural sciences.be/belgica/en/index.

Data processing was performed at the RCMG by means of the software SonarScope. Processing consisted of statistical spike removal, manual cleaning of artefacts and tide correction. No special filtering was necessary due to the good quality of the raw data (Figure 3.8). Tide corrections were carried out by applying information provided by the French Hydrological service (SHOM) on the water elevation at the different times and locations we surveyed. The corrected bathymetric set was gridded at 5 m bin size. The final bathymetric DTM exhibits minor local elevation mismatches between consecutive swaths due to inaccuracies in the tidal model. These errors could not be removed completely. Fortunately, they distinguishable from real seabed features.

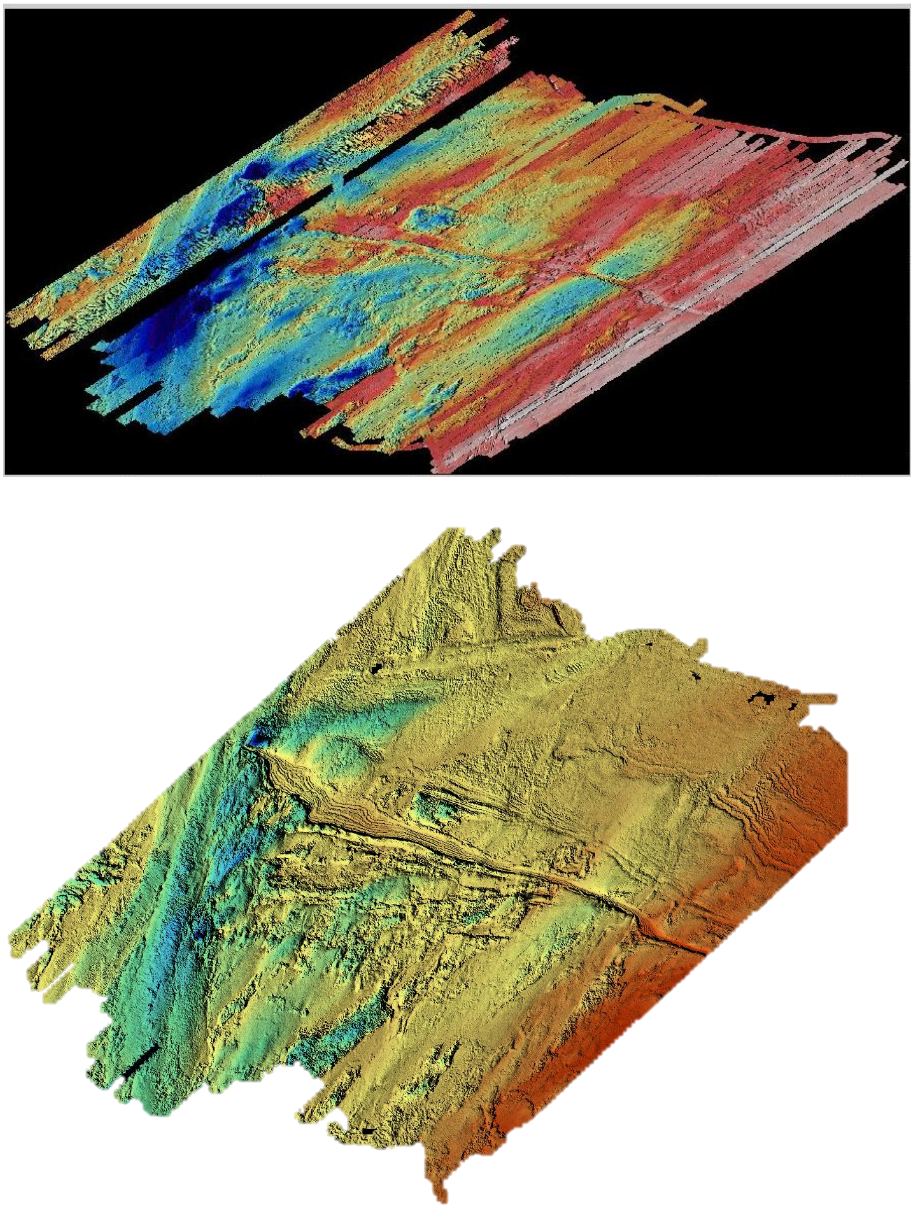

Figure 3.8. Digital Terrain Model derived from ROB–RCMG multibeam bathymetry. Above, raw data; below: processed data.

3.2.1.2. MCA–UKHO bathymetric dataset

This dataset was acquired under contract to the British Maritime and Coastguard Agency (MCA) as part of their “Civil Hydrographic Program”. These data were collected along the HMOI 1159 Dover Strait Traffic Separation Scheme.

Bathymetric surveys were undertaken by the UK Hydrographic Office (UKHO) in 2006 and 2007 on board of different vessels and using a variety of different survey systems. In total, three vessels were employed: M/V Victor Hensen, M/V Geniusbank, and S/V Meridian. Data from M/V Victor Hensen represents the main set of lines. This vessel was equipped with a

Kongsberg EM 710 multi-beam echosounder. This equipment operates with frequencies ranging from 60 kHz to 100 kHz, a maximum ping rate of 25 Hz and 400 beams equiangular spaced in target detection mode. The M/V Meridian was equipped with MBES Reson SeaBat 8101ER and the SeaBat 7125 with operating frequencies of respectively 240 kHz and 400 kHz, while a MBES Kongsberg EM 3002D (operating frequency of 300 kHz) was on board M/V Geniusbank.

This dataset was provided to Imperial College London as pre-cleaned and tide-corrected along-track soundings in GSF file format for research purposes. It was gridded at Imperial College London at 1.5 m bin size by using DMagic Fledermaus v.7. XYZ text files and grids in ArcGrids format of the latter were provided to the author of the present study. Detailed information on this dataset can be consulted in Oggioni (2013). This dataset also presents minor linear artefacts. These were easily identifiable, for they are parallel to the ship survey-tracks and show different patterns when compared to real seabed linear features identified in other areas of the strait.

3.2.1.3 UKHO English Channel grid

This dataset consists of historical single-beam echo soundings (SBES) of the UK sector of the English Channel collected by the Maritime and Coastguard Agency (MCA) between 1988 and 2004 and archived at the UK Hydrographic Office (UKHO). These data were provided to our colleagues of Imperial College London merged in a 30 m cell-size grid. They were furnished to the author of this manuscript as XYZ text files and 30 m cell-size grids in ArcGrids format. Detailed information on this dataset can be consulted in Oggioni (2013).

3.2.1.4 SHOM single-beam bathymetric data

These data consist of a collection of single-beam datasets collected by the Hydrographic and oceanographic service of the French Marine (SHOM) during the past forty years. This dataset covers the French sector of the Dover Strait. These soundings were gridded at Lille 1 University (France) by using IV3D DMagic and Fledermaus v.7 under the terms of contract N° 205/2011. The density of the soundings in the dataset was highly variable, ranging from areas with high or complete coverage to zero or very poor coverage. Therefore, two grids were produced: one at 80 m and another at 40 m cell size. These data were provided to the author of this dissertation as XYZ text files and in ArcGrid format. Detailed information on this dataset can be consulted in Oggioni (2013).

3.2.2. Southern North Sea bathymetric dataset

The southern North Sea bathymetric data comprise 5 datasets (Figure 3.6); i.e. The BCS dataset, the DCS dataset, the GSB dataset, the OTE REC dataset, and the EC REC dataset. These data were complemented with a 40 m bin-size bathymetric grid merged from the various sets of data composing the Dover Strait bathymetric dataset. These datasets have been completed with Emodnet bathymetric data in order to cover the entire study area.

The Belgian Continental Shelf (BCS) bathymetric dataset is a compilation of single-beam and multibeam bathymetry acquired between the 1990s and 2015 by the Belgian Flemish Hydrography and the Belgian FPS Economy Continental Shelf Services. This dataset was provided as ArcGrid files by the Belgian Flemish Hydrography Service fully processed and gridded at 20 m bin size.

The Dutch Continental Shelf (DCS) bathymetric dataset was provided by Deltares (data available at www.deltares.nl). It comprises the single-beam and multibeam bathymetric data gathered by Deltares until 2013. Deltares furnished this dataset fully processed and gridded at 25 m bin size.

The Gabbard Sandbanks (GSB) bathymetric dataset was collected in 2005 by the Greater Gabbard Offshore Wind Consortium in the framework of the Greater Gabbard windfarm site characterization. These data were downloaded from www.marinedataexchange.co.uk fully processed and gridded at 10 m bin-size. Reports on acquisition and processing are also available at that website.

The Outer Thames Estuary (OTE) and East Coast (EC) Regional Environmental Characterizations (REC) bathymetric grids were downloaded from www.marinedataexchange.co.uk; reports on acquisition and processing are also available at that website. These datasets consist of 1 m bin-size bathymetric DTMs, which were provided as XYZ text files processed and tide-corrected. These grids were derived from the combination of seismic reflection profiles, multibeam bathymetry and side-scan sonar data acquired along ~400 m wide bands spaced 2.5–6 km and 7–13 km, and acquired along East Anglia and the Outer Thames Estuary, respectively (Figure 3.6). The East Coast REC dataset was acquired and processed by Limpenny et al. (2011). The OTE datasets were acquired in 2008 by Emu Ltd and the Institute of Estuarine Coastal Studies of University of Hull and processed and interpreted by Emu Ltd & University of Southampton (2009). Backscatter GeoTiffs are also available along those bands. The 400 m wide bands along which these data were collected are too far apart from one another to grid them together in a single DTM. In this study, we used these datasets to better visualize minor features and calculate heights and depths more accurately. Otherwise, we base our analysis on the other less accurate, but more complete, bathymetric datasets.

References

De Clercq, M., ChaDTMenos, V., Van Lancker, V., and Missiaen, T., 2016, A high-resolution DTM for the Top-Paleogene surface of the Belgian Continental Shelf, Journal of Maps, v. 12, p. 1047–1054.

Depret-G-tec, 2009, Seismisch onderzoek in Exploratiezone 4 op het Belgisch Continentaal Plat.

Hughes-Clarke, J.E., Mayer, L.A., and Wells, D.E., 1996, Shallow-water imaging multibeam sonars: A new tool for investigating sea floor processes in the coastal zone and on the continental shelf, Marine Geophysical Researches, v. 18, p. 607–629.

08D-005-Depret-SeisVibro/MA/GP/RE001, 236 p.

Kint, L. and Van Lancker, V., 2016, SediLITHO@SEA v2 (06/10/2016). Database lithological descriptions, with relevance to Belgian part of the North Sea, Brussels: Royal Belgian Institute of Natural Sciences.

Limpenny, S.E., Barrio Froján, C., Cotterill, C., Foster-Smith, R.L., Pearce, B., Tizzard, L., Limpenny, D.L., Long, D., Walmsley, S., Kirby, S., Baker, K., Meadows, W.J., Rees, J., Hill, J., Wilson, C., Leivers, M., Churchley, S., Russell, J., Birchenough, A.C., Green, S.L., and Law, R.J., 2011, The East Coast Regional Environmental Characterisation, Cefas Open report 08/04 (available at www.alsf-mepf.org.uk), 287 p.

Mathys, M., 2009a, The Quaternary geological evolution of the Belgian Continental Shelf, southern North Sea, PhD thesis, Ghent University, Belgium, 454 p.

Mathys, M., Van Lancker, V., Versteeg, W., and De Batist, M., 2009b, Wetenschappelijke begeleiding en geïntegreerde interpretatie van seismisch onderzoek en trilboringen in Exploratiezone 4 op het Belgisch Continentaal Plat. Rapport Vlaamse Overheid, Agentschap voor Maritieme Dienstverlengingen Kust, Afdeling Kust, 146 p.

Oggioni, F., 2013, Characterization of catastrophic flood-related features in the English Channel. PhD thesis, Imperial College London, 330 p.

Stockwell, J.W., Jr., 1999, The CWP/SU: Seismic Unix package, Computers and Geosciences, v. 25, p. 415-419.

Vanneste, K., Verbeeck, K., Garcia-Moreno, D., Versteeg, W., and Jomard, H., 2011, Results of the seismic reflection survey Belgica 2010/09 “North-Artois Shear Zone”, Technical Report, 16 p.

Yilmaz, Ö., 2001, Seismic data analysis: Processing, inversion, and interpretation of seismic data, Society of exploration geophysicists, 2065 p.

Chapter 4

Fault activity in the epicentral area of the 1580 Dover Strait (*Pas-de-Calais*) earthquake (northwestern Europe)

Chapter 4 – Fault activity in the epicentral area of the 1580 Dover Strait (Pas-de-Calais) earthquake (northwestern Europe)

This chapter is a slightly modified version of the manuscript that was published as:

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Abstract

On 1580 April 6 one of the most destructive earthquakes of northwestern Europe took place in the Dover Strait (*Pas de Calais*). The epicenter of this seismic event, the magnitude of which is estimated to have been about 6.0, has been located in the offshore continuation of the North Artois Shear Zone, a major Variscan tectonic structure that traverses the Dover Strait. The location of this and two other moderate magnitude historical earthquakes in the Dover Strait suggests that the North Artois Shear Zone or some of its fault segments may be currently active. In order to investigate the possible fault activity in the epicentral area of the AD 1580 earthquake, we have gathered a large set of bathymetric and seismic reflection data covering the almost-entire width of the Dover Strait. These data have revealed a broad structural zone comprising several subparallel WNW–ESE trending faults and folds, some of them significantly offsetting the Cretaceous bedrock. The geophysical investigation has also shown some indication of possible Quaternary fault activity. However, this activity only appears to have affected the lowermost layers of the sediment infilling Middle Pleistocene palaeo-depressions. This indicates that, if these faults have been active since Middle Pleistocene, their slip rates must have been very low. Hence, the AD 1580 earthquake appears to be a very infrequent event in the Dover Strait, representing a good example of the moderate magnitude earthquakes that sometimes occur in plate interiors on faults with unknown historical seismicity.

Keywords: Palaeoseismology; Intraplate processes; Submarine tectonics; Neotectonics; Europe.

Author contributions: T.C. and M.D.B. conceived, designed and coordinated the study. D.G-M., K.VAN. (K. Vanneste) and W.V. designed, coordinated and conducted RV Belgica geophysical surveys with important contributions from K.VER. (K. Verbeeck) and H.J. K.VAN. and O.Z.H. processed RCMG–ROB and Dangeard I seismic reflection data. D.G-M. processed RCMG–ROB bathymetric data. F.O. and J.S.C. processed MCA–UKHO and MCA bathymetric data. J.S.C., F.O. and S.G. provided MCA and MCA–UKHO data. A.T. provided Dangeard I seismic reflection data and SHOM bathymetry. H.J. provided borehole information. D.G-M analyzed and interpreted the data with contributions from F.O., K.VAN. and K.VER. All authors discussed the results. D.G-M wrote the paper with important contributions from K.VAN. and K.VER. All authors reviewed the paper before submission.

4.1 Introduction

Assessing the seismic hazard associated with plate-interior tectonic structures is generally a very difficult task. For instance, the seismicity in these areas is generally too low to be studied by means of classic seismotectonic approaches. In addition, destructive earthquakes are rare and they usually appear randomly distributed in space and are sometimes located in places with no known historical seismicity (Zoback and Grollimund, 2001; Camelbeeck et al., 2007). This apparent randomness of the seismicity might be the consequence of recurrence intervals too long to have produced two significant ruptures of the same feature in historical times (Stein and Mazzotti 2007). It is thus necessary to adopt palaeoseismic methods in order to understand and characterize the activity of intraplate faults and hence the possible present-day seismic hazard related to them.

This study focuses on the poorly known offshore Sangatte Fault, which traverses the marine Dover Strait (Pas de Calais) from Sangatte (northeastern France) to Folkestone (southeastern England; Figure 4.1 and Figure 4.2). This fault is part of the North Artois Shear Zone (Figure 4.1) and it is believed to be the probable source of the magnitude ∼6.0 earthquake that occurred offshore in AD 1580 (Camelbeeck et al., 2007). Historical information suggests that this earthquake produced damage equivalent to MSK intensity VIII in Calais and Dover and about VI in London, and it was felt as far as Köln (Germany) to the east and York (England) to the north (Neilson et al., 1984; Melville et al., 1996; Musson, 2004). According to the empirical relationships of Wells and Coppersmith (1994), the inferred magnitude of the AD 1580 earthquake would correspond to an average fault length of 10 km and to a slip of 0.25–0.30 m. However, nothing is currently known about any active fault located in the Dover Strait.

Apart from the AD 1580 earthquake, the Sangatte fault has only been associated with two other seismic events of moderate magnitude. These are the MS ∼4.0 historical earthquake that occurred in 1776 (Melville et al., 1996) and the Mw 4.0 Folkestone earthquake of 2007 (Ottemöller et al., 2009). The hypocenter of the latter was instrumentally localized about 5±2 km below Folkestone (Ottemöller et al., 2009).

The presence of major infrastructures and the densely populated cities within the area of influence of an earthquake similar to the AD 1580 event make it very important to assess the nature of the Sangatte Fault and its tectonic activity over time. In order to do so, we have gathered a large set of seismic reflection and bathymetric data from the epicentral area of this earthquake. In this paper, we will present and discuss the interpretation of these geophysical data, which resulted in the first high-resolution bathymetric map of the Dover Strait and provided a clear visualization of the main tectonic structures and their activity over time.

4.2 Tectonic and geological settings

The North Artois Shear Zone is a complex fault-and-fold system defining the western part of the Variscan Midi–Eifel Thrust Front (Figure 4.1), which separates the Ardenne Massif and Paris Basin to the south (hanging wall) from the London–Brabant Massif to the north (footwall). The present-day geometry of this fault zone results from several phases of post-Paleozoic deformation that have induced different reactivations of structures inherited from the Variscan Orogeny (e.g. Chadwick et al., 1993; Vliet-Lanoë et al., 1998; Vliet-Lanoë et al., 2002a,b; Mansy et al., 2003; Minguely et al., 2010). Since its formation, the North Artois Shear Zone has passed from the original compressional setting (Variscan Orogeny) through extension related to the opening of the Tethyan and Atlantic ocean basins during Jurassic and Cretaceous times, and again to compression during the tectonic inversion that started in the early Paleogene epoch due to the Alpine Orogeny (Vandycke et al., 1988; Bergerat and Vandycke 1994; Van Vliet-Lanoë et al., 2002a,b; Van Vliet-Lanoë et al., 2004; Mansy et al., 2003; Minguely et al., 2010).

From an analysis of gravity data, Everaerts and Mansy (2001) concluded that the major fault segments comprising the North Artois Shear Zone have lengths ranging from 15 to 40 km and are arranged as right-stepping en-echelon fault zones (Figure 4.2). Based on their gravity map, Camelbeeck et al., (2007) distinguished five major faults; these are: Marqueffles Fault, Ruitz Fault, Pernes Fault, Landrethun Fault and Sangatte Fault (see Figure 4.2). Not only their activity, but the precise geometry of these faults is still poorly known. This is especially true for the mostly submarine Sangatte Fault, on which there has been no specific study published until now; despite the fact that it corresponds to the most likely geological structure capable of having generated the AD 1580 event (e.g. Camelbeeck et al., 2007).

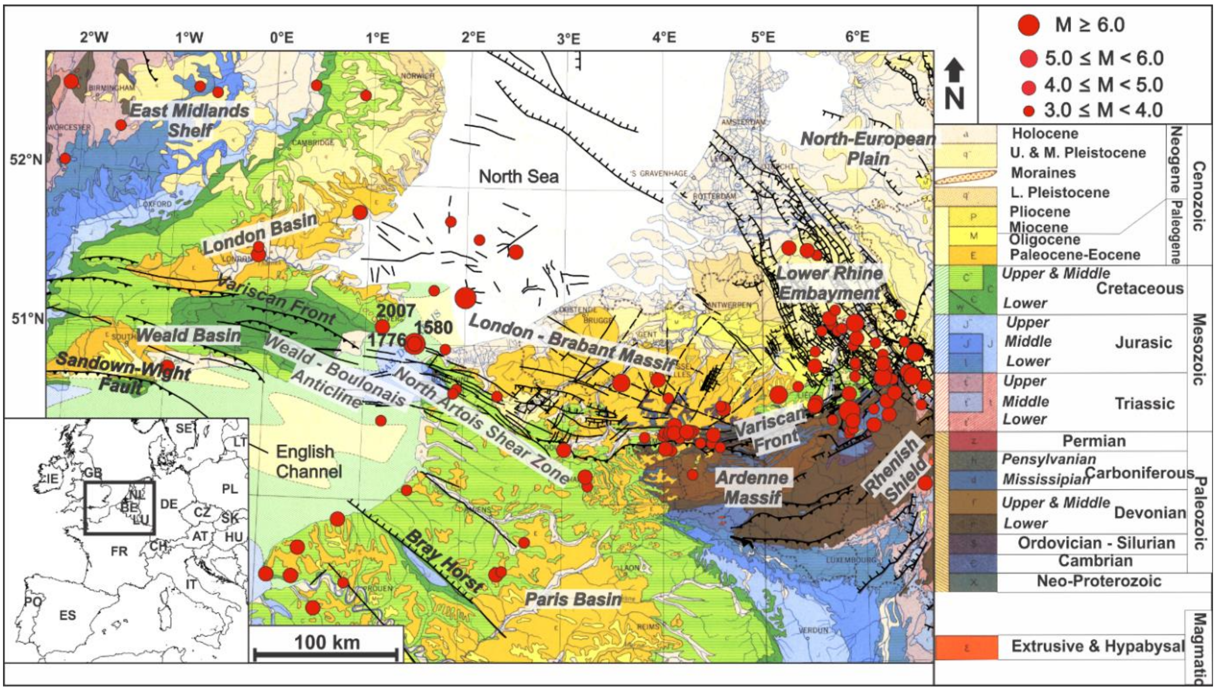

Figure 4.1. Onshore geological map of northwestern Europe updated with the geology of the English Channel and major tectonic structures (De Béthune and Bouckaert 1968). Known historical and instrumental earthquakes (Mw ≥ 3) are indicated scaled by magnitude according to catalogues from the Royal Observatory of Belgium and Grünthal et al. (2009).

The Quaternary activity of the North Artois Shear Zone is still debated, as no conclusive field evidence of recent tectonic deformation has yet been associated with any of its fault segments. Indirect evidence includes possible extensional faults identified in the Sangatte cliff (Van Vliet-Lanoë et al., 2000) and minor right-lateral deformations affecting river development and Quaternary fluvial and aeolian deposits in northeastern France (Colbeaux et al., 1981; Van Vliet-Lanoë et al., 2002a; Mansy et al., 2003). The latter suggest right-lateral strike-slip deformation, which is in good agreement with the NNW–SSE orientation of the maximum horizontal stress measured near Boulogne by Froidevaux et al. (1980). However, it disagrees with the direction of the deformation suggested by the focal mechanism calculated for the 2007 Folkestone earthquake (Ottemöller et al., 2009), which indicates a left-lateral rupture along a WNW–SES fault (see Ottemöller et al., 2009).

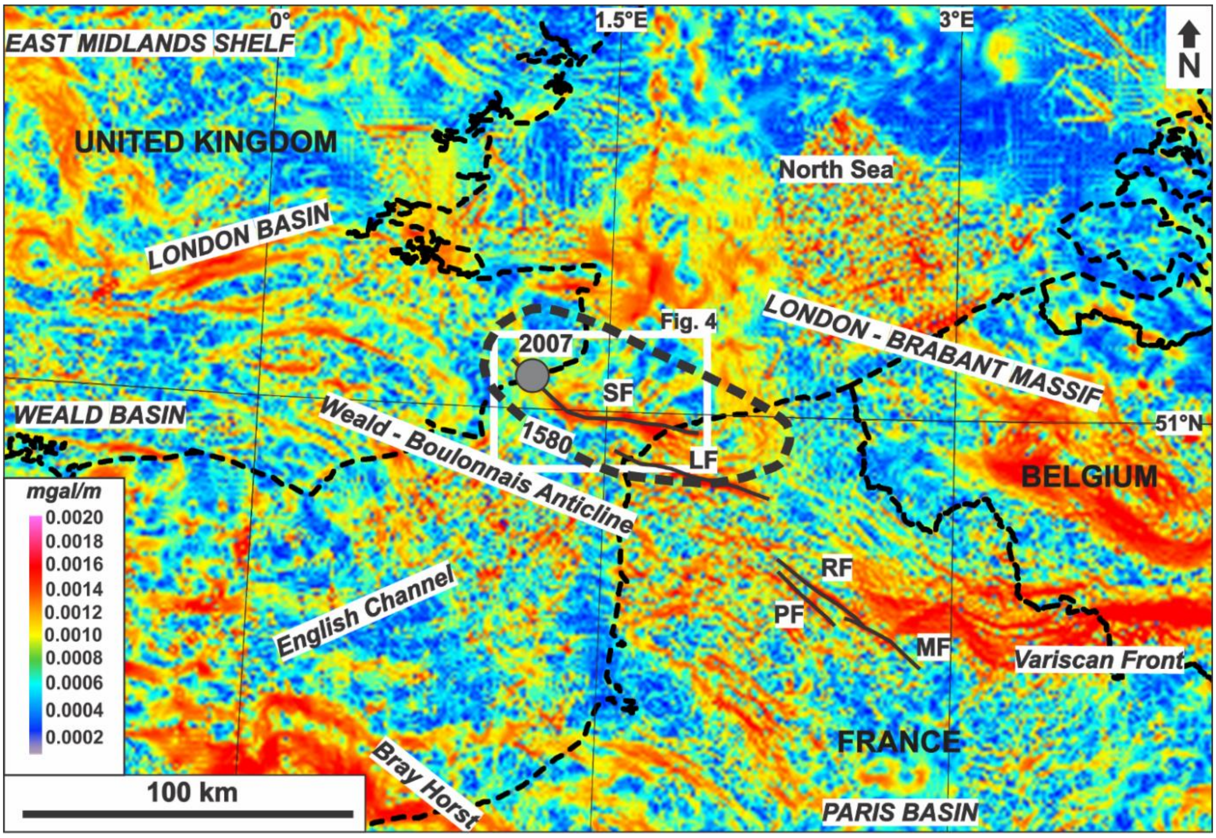

Figure 4.2. Interpreted horizontal derivative of the Bouguer gravity anomaly (see Everaerts and Mansy 2001; Camelbeeck et al., 2007). The Bouguer anomaly was calculated using the gravity database of the Royal Observatory of Belgium. For France and United Kingdom the data were provided by the French and British Geological Surveys respectively. All gravity data are referenced to the gravity datum Uccle (1976) (IGSN71–0.048 mGal). The density reduction used for the calculation of the Bouguer anomaly on land was 2.67. Above the sea, the free-air anomaly was used. Grey colored circle: epicenter of the 2007 Folkestone earthquake (Ottemöller et al., 2009); dashed ellipsoid: isoseismal indicating MSK Intensity VII–VIII during the 1580 earthquake (Melville et al., 1996); LF, Landrethun Fault; SF, Sangatte fault; RF, Ruitz fault; PF, Pernes fault and MF, Marqueffles fault. White rectangle: area shown in Figure 4.4.

According to some authors, the Sangatte Fault may have been in active extension during the deposition of the Gault Clay and Lower Chalk Formations in the Cretaceous (Warren and Harris 1996; Ottemöller et al. 2009; Minguely et al., 2010). On the 1:250 000 offshore geological maps published in 1988 and 1989 by the British Geological Survey (BGS) and the compilation of geological data performed by James et al. (2002), the Sangatte Fault is represented as several WNW–ESE trending minor faults and folds traversing the Cretaceous bedrock (Figure 4.3). This structure also traverses two significant Quaternary features situated in the center of the Dover Strait (Figure 4.3 and Figure 4.4): a broad palaeo-channel known as the “Lobourg Channel”, and a complex sediment-filled palaeo-depression network known as the “Fosses Dangeard” (see Destombes et al., 1975; James et al. al., 2002). Elsewhere Quaternary sedimentary features are mainly limited to a number of major (stable) and minor (mobile) Holocene sandbanks separated by areas with thin or absent sedimentary cover (see James et al., 2002). Recent tectonic movements affecting the latter features will not be preserved for long due to the dynamic conditions this area has been subjected to during the Holocene, with strong sediment reworking, erosion and sediment starvation (e.g. Hamblin et al., 1992). On the other hand, if the Sangatte Fault reaches the surface and has ruptured several times in earthquakes similar to the AD 1580 since the Fosses Dangeard and the Lobourg Channel were formed, the cumulated deformation produced on these erosional features should be measurable.

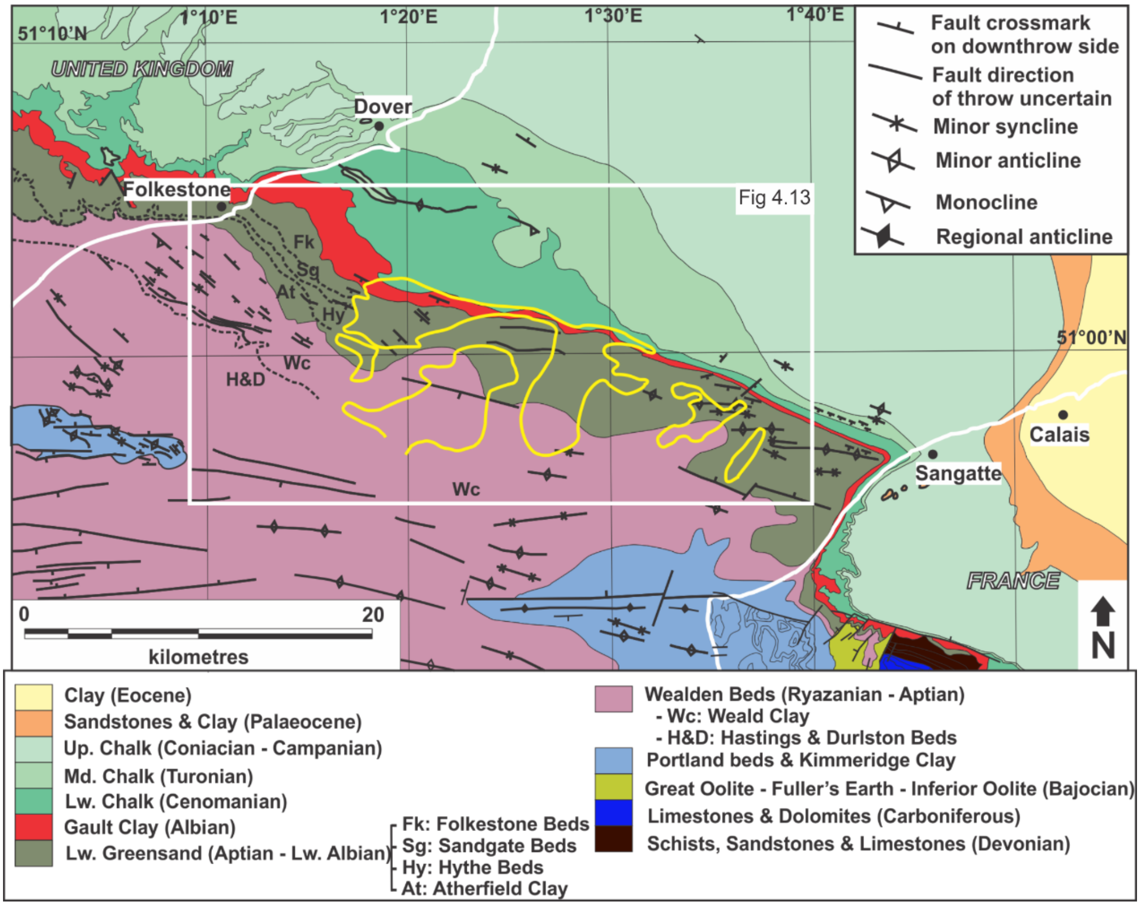

Figure 4.3. Bedrock geology of the Dover Strait/Pas-de-Calais area constrained by combining the 1:250 000 geological maps published by the BGS in 1988 and 1989 (sheets 51N00 and 50N00) and the maps published by James et al. (2002). White rectangle indicates the area shown in Figure 4.13. The yellow line represents the location of the Fosses Dangeard according to James et al. (2002).

The Fosses Dangeard are a kilometer-scale, interconnected set of sediment-filled palaeo-depressions that are incised several tens of meters into bedrock (e.g. Destombes et al., 1975; Smith 1985; James et al., 2002). The origin of these buried palaeo-depressions is unknown, although most authors agree that they were probably formed during Middle Pleistocene (e.g. Gibbard, 1995; Gibbard and Cohen, 2015). Present consensus holds that the Fosses Dangeard were carved during the Elsterian glacial maximum (0.45 Ma BP) due to water overspilling the Weald-Artois ridge, which used to connect northern France with southeastern England (e.g. Smith, 1985; Hamblin et al., 1992; Gibbard, 1995; Gibbard and Cohen, 2015). According to palaeogeographic reconstructions, the blockage of the northern routes of the southern North Sea drainage systems by the coalescence of the ice sheets across the North Sea during that glacial maximum induced the formation of a lake in the unglaciated area of the southern North Sea (e.g. Gibbard, 1995; Gibbard and Cohen, 2015). That lake was dammed in the southwest at the Weald–Artois ridge, which, eventually, was overtopped by the rising lake level and evolved into a waterfall (Gibbard and Cohen, 2015). That ridge appears to have breached eventually, ending the waterfall phase and initiating the opening of the Dover Strait (e.g. Gibbard and Cohen, 2015). Hamblin et al. (1992) and Mellett et al. (2013), among others, proposed that the breach (i.e. the opening of the Dover Strait) and the incision of the Fosses Dangeard are the result of progressive fluvial erosion, which also incised part of the Channel palaeovalleys. Alternatively, Smith (1985), Gupta et al. (2007) and Collier et al. (2015) have proposed that the Fosses Dangeard formed by plunge-pool erosion followed by catastrophic flooding (also known as megaflood), which carved some of the Channel palaeovalleys. According to these authors, that megaflood was generated by a sudden breach of the ridge damming the lake at the Dover Strait.

The hypotheses exposed above on the opening of the Dover Strait and the incision of the Fosses Dangeard are not the only ones. For instance, Van Vliet-Lanoë et al. (2004) suggested a possible tectonic origin/control of the erosion. According to these authors, the Dover Strait went through a series of closing-and-opening episodes during the Quaternary Period, which was induced by a reactivation of the North Artois Shear Zone by a combination of glacio-isostatic adjustments with background tectonic movements. However, a tectonic origin of these features is very unlikely, as considering the amount of deformation necessary to create such large depressions, similar offsets should be observable in Quaternary deposits along the on-shore continuation of these structures, which is not the case (e.g. Minguely et al., 2010). However, it is indeed possible that this fault system and/or its possible Quaternary tectonic activity might have had some control on the erosion of the Fosses Dangeard and Lobourg

Channel.

Absolute dating on the sediments infilling of the Fosses Dangeard is limited to pollen samples extracted from a 50 m borehole collected from one of its sediment-filled palaeo-depressions (Destombes et al., 1975). This dating suggests that the sampled sediments were deposited during the Brørup interstadial of the Weichselian glaciation. Destombes et al. (1975) argued that the formation of the Fosses Dangeard should however be much older, as the features themselves reach depths up to 100 m. In addition, the sedimentary infill of these palaeo-depressions presents several internal erosional surfaces, attesting to different phases of scouring and infilling. As aforementioned, it is indeed generally accepted (e.g. Gibbard 1995; Toucanne et al., 2009) that the incision of these features occurred during the Elsterian glaciation (0.48–0.40 Ma BP). The age of the sediment infilling the Fosses Dangeard might therefore comprise sediments deposited from the Elsterian glacial maximum (0.45 Ma BP) up to the Weichselian glaciation (0.11–0.012 Ma BP).

The Lobourg Channel is a broad NE–SW palaeo-channel extending across the Dover Strait and cut predominantly into bedrock (James et al., 2002). This palaeo-channel extends to the southwest into a complex anastomosing system of valleys (i.e. The Channel palaeovalley system), which can be mapped continuously from a few kilometers northeastward of the Dover Strait to the western approaches (e.g. Lericolais et al., 2003; Mellett et al., 2013). Currently, two main hypotheses are proposed to explain this major erosional feature. Some authors (e.g. Gibbard 1995; Lericolais et al., 2003) suggest that these palaeovalleys were part of a major palaeo-drainage system (i.e. the Channel/Manche River) that was fed by northwestern European Rivers during major Middle and Late Pleistocene marine lowstands. Others authors have proposed that this feature may have been created by several episodes of Middle Pleistocene catastrophic flooding (e.g. Smith, 1985; Gupta et al., 2007; Collier et al., 2015). In any case, both groups of authors agree that the Lobourg Channel formed following the breach of the Weald-Artois ridge. This palaeo-channel appears to have hosted a river during all major Middle–Late Pleistocene marine lowstands, being finally submerged at the beginning of the Holocene epoch (e.g. Gibbard, 1995; Lericolais et al., 2003; Toucanne et al., 2009). Therefore, the age of the erosion that shaped the Lobourg Channel ranges between 0.45 Ma and 0.012 Ma.

4.3 Methodology and available data

This study is based on the interpretation of several sets of bathymetric and seismic reflection data (Figure 4.4) collected across the gravity anomaly observed by Everaerts and Mansy (2001) traversing the Dover Strait (see Figure 4.2). The principal seismic reflection dataset comes from three geophysical campaigns organized by the Royal Observatory of Belgium (ROB) in collaboration with the Renard Center of Marine Geology (RCMG) in 2010 and 2012 on board of RV Belgica (cruise reports 2010/09, 2012/03 and 2012/25 available at www.mumm.ac.be). These campaigns consisted of the acquisition of 17 multichannel (total 309 line km) and 48 single-channel (total 487 line-km) seismic reflection profiles over the entire width of the strait and more than 300 km² of multibeam bathymetric data from its central part. These data were complemented with 20 other high-resolution seismic-reflection profiles acquired from the central part of the Dover Strait by the RCMG and Lille University in 2002 on board of RV Sepia II (Dangeard I Cruise), as well as with parts of several bathymetric datasets gathered by Imperial College London. Technical details on the vertical and horizontal resolutions of the geophysical data, as well as on their acquisition and processing, can be consulted in "Chapter 3 – Methods" of this dissertation. Location and coverage of the various geophysical surveys are shown in Figure 3.3.

The seismic dataset mainly consists of parallel lines. This was due to limitations of collecting lines across the Dover Strait due to strong tides and shipping lane restrictions. The high quality of the bathymetric data was therefore invaluable to correlate between the seismic profiles. Notably, the high resolution of the bathymetric datasets acquired from the center of the Strait (horizontal resolution 1.5–5 m) and that of the single-channel seismic-reflection (vertical resolution 1–3 m) datasets permitted the visualization of features as small as 1–1.5 m in the center of the strait.

The processed seismic data were imported into the interpretation software Kingdom Suite (TKS) and Opendtect (http://www.opendtect.org). Stratigraphic units and Quaternary erosional surfaces (when possible) were correlated from one profile to another by using the three-dimensional software Opendtect.

The final bathymetric merge was produced using QPS Fledermaus DMagic and Global Mapper 14. Individual Arc Grid files (.asc) from the previously gridded .sd files were used for the 'scene file merge'. The Arc Grid files were loaded into Global Mapper in a single workspace and overlapped in order to display the data set at the highest resolution in each area. The workspace comprising all the different grids was then saved into a single Arc Grid file respecting the different cell size of each of the grids composing the map (Figure 4.4).

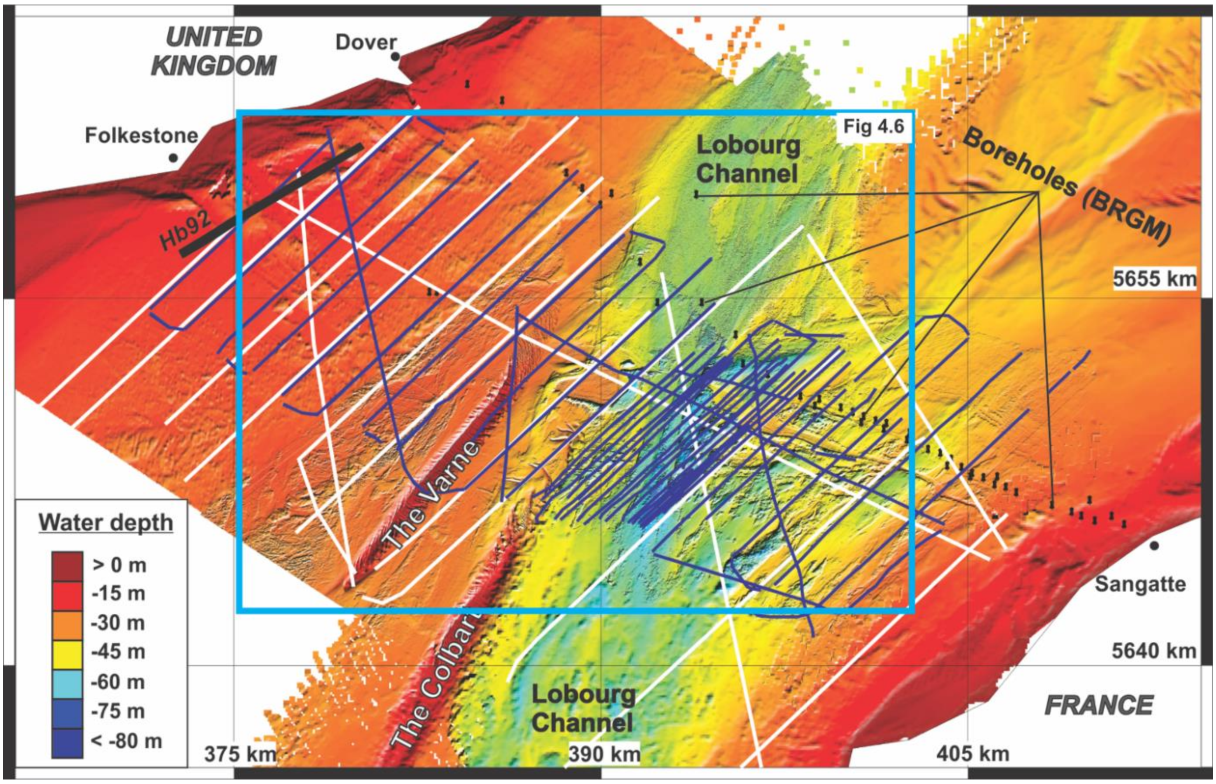

Figure 4.4. Merged bathymetry, single-channel (dark blue lines) and multichannel (white lines) seismic reflection profiles gathered for this study, and boreholes archived at BRGM. Hb92: location of the single-channel seismic-reflection profile interpreted by Hamblin et al. (1992). Light blue rectangle: area shown in Figure 4.6. Projection for this and following figures: UTM-31N (WSG84); distances are given in kilometers.

4.4 Interpretation of the Geophysical data

4.4.1 Seismic stratigraphic interpretation

In this study, correlations between geology and seismic stratigraphy were first done in the northwesternmost seismic reflection profile collected for this study and correlated afterwards southeastward from one profile to another. The interpretation of the first profile is based on the equivalence established by Hamblin et al. (1992) between geology and the seismic stratigraphy of one single-channel seismic-reflection profile acquired nearby the English shore (see location in Figure 4.4). We complemented this interpretation with several interpreted boreholes (Figure 4.4) made available by the BRGM (www.brgm.fr) and by comparing with the currently available geological maps from the Dover Strait (Figure 4.3). Correlations between seismic profiles were performed by combining the few available cross-lines with the geomorphological expression of the various geological units in the bathymetry. The latter proved invaluable in correlating between the seismic profiles, as many of the sedimentary formations form distinct geomorphological features on the seafloor that can be followed along the entire width of the strait (see Figure 4.5, Figure 4.6 and Figure 4.8). The geological interpretation presented in this study is limited to the first 100 m below the seabed, as we were not able to completely remove the first seabed multiple in the multichannel seismic-reflection data. Hence, we have very little information on the structure below the Wealden Beds of Lower Cretaceous age (Figure 4.7).

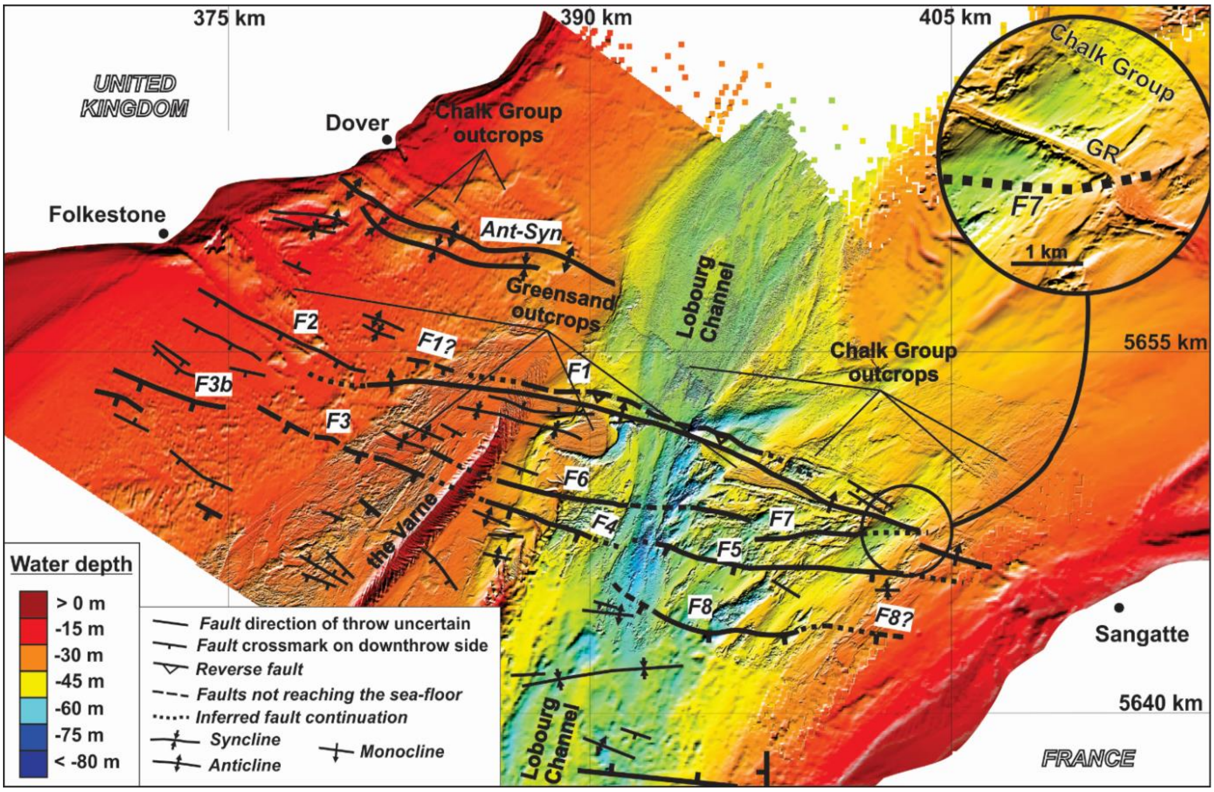

Figure 4.5. Structural map derived from the interpretation of the seismic reflection and multibeam bathymetric datasets collected for this study. Major bathymetric features are indicated. Ant-Syn: Anticline/Syncline system; GR: Greensand Ridge.

The strata imaged in the seismic reflection profiles correspond to Lower and Upper Cretaceous sedimentary rocks, as well as sediments infilling Quaternary palaeo-depressions and palaeo-channels or forming sandbanks. Paleogene and Neogene units are missing in this area. The Quaternary palaeo-depressions and palaeo-channels are the easiest features to identify in the seismic reflection data due to the distinct unconformity formed by their basal erosional surfaces (BES). The seismic facies infilling the largest palaeo-depressions (i.e. the Fosses Dangeard) comprise several subfacies composed of seismic reflections with low to moderate amplitude. These subfacies are separated (and truncated) by thin packages of high amplitude reflections that seems to correspond to erosional surfaces carved into the sedimentary infill (Figure 4.9).

During this study, we mapped the infill of Quaternary erosional features (e.g. the Fosses Dangeard), constructing isopach maps along the different seismic reflection profiles by subtracting the two-way times (TWT) of their BES and seabed. The resulted thickness map was converted to depth assuming an average velocity of 1800 m s^{-1}, as proposed by Arthur et al. (1997) from log tests performed for the Channel Tunnel geo-engineering investigations. We have included the isopach map in Figure 4.6 to illustrate the distribution of the Quaternary deposits with regards to the main tectonic structures.

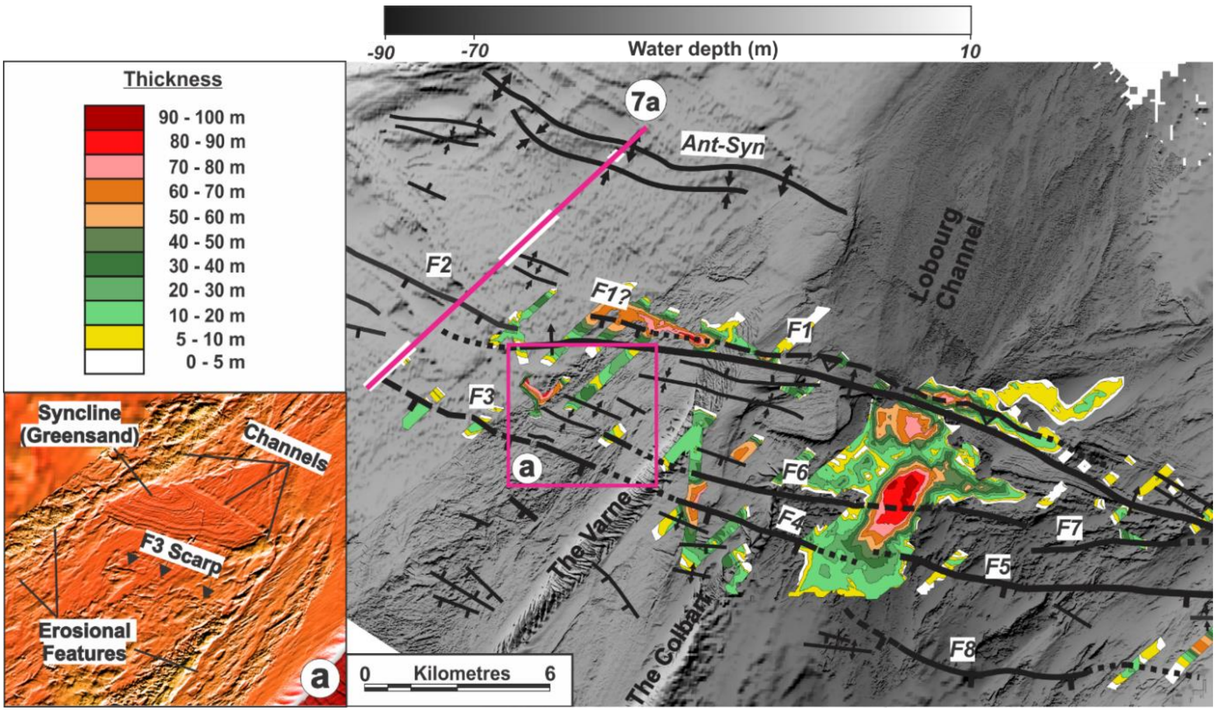

Figure 4.6. Structural interpretation and isopach map of the sediments infilling Quaternary palaeo-depressions and palaeo-channels (see Figure 4.4 for location). (a) Enlarged view of the purple square indicated in the structural map. Location of the single-channel seismic-reflection profile shown in Figure 4.7(a) is indicated.

Concerning the seismic-stratigraphy of the Cretaceous bedrock, we identified the following seismic units, from Lower to Upper Cretaceous age:

(1) The upper part of the fluvial and lacustrine Wealden Beds (see Hamblin et al., 1992;

Radley and Allen, 2012), which are characterized by a succession of well-defined low and high amplitude reflections in its upper part that become discontinuous and heterogeneous with depth. The seismic reflection profiles show significant lateral variability of this unit's seismic facies, with the presence of buried palaeo-channels and other erosional/depositional features consistent with its continental origin (Figure 4.7b).

(2) The shallow marine/shoreline deposits known as the Lower Greensand (see Ruffell and Wach 1991; Hamblin et al., 1992; Hopson et al., 2010), which are divided into four units in the eastern English Channel (e.g. Hamblin et al., 1992; Hopson et al., 2010); from old to young: Atherfield Clay, Hythe Beds, Sandgate Beds and Folkestone Beds. The Atherfield Clay presents an almost completely acoustically transparent facies (Figure 4.7a). The Hythe Beds appear as a package of reflections of moderate amplitude. The Sandgate Beds consist of seismic subunits composed of reflections of moderate amplitude alternating with seismic subunits comprising reflections with relatively lower amplitudes (Figure 4.7a). Finally, the Folkestone Beds are characterized by very diffuse and wavy reflections presenting hyperbolic diffractions (Figure 4.7a). In general, the Lower Greensand Formation thins to the southeast. This is especially true for the upper part of the sequence, with the Sandgate Beds thinning by as much as 25 per cent and the Folkestone Beds being no longer recognizable in the southeastern half of the strait (Figure 4.8). Assuming that the Folkestone Beds are absent in the southeast (see next point), the Lower Greensand Formation is ~30 m thicker offshore England than offshore France (compare Figure 4.7 and Figure 4.8).

(3) The marine deposits known as Gault Clay (see Hamblin et al., 1992; Hopson et al., 2010) present two distinct parts. The upper half is acoustically almost transparent, whereas the lower half has a more heterogeneous acoustic character with the appearance of low amplitude reflections (Figure 4.7). These two subunits appear to be equivalent to the Upper and Lower Gault Clay described by Owen (1975) and Woods et al. (1995). This subdivision is clearly observed near the British coast. However, it becomes more ambiguous towards France, where a seismic unit showing very similar characteristics to the lower Gault Clay lies directly on top of a seismic unit resembling the Sandgate Beds (Figure 4.8 and Figure 4.9). The Folkestone Beds appear thus to be missing in that area.

The thickness of the Gault Clay Formation ranges between 30 and 50 m across the survey area, generally getting thinner to the southeast. This is consistent with its outcrop in northern France, where only 11 m of this unit are recorded (e.g. Hamblin et al., 1992).

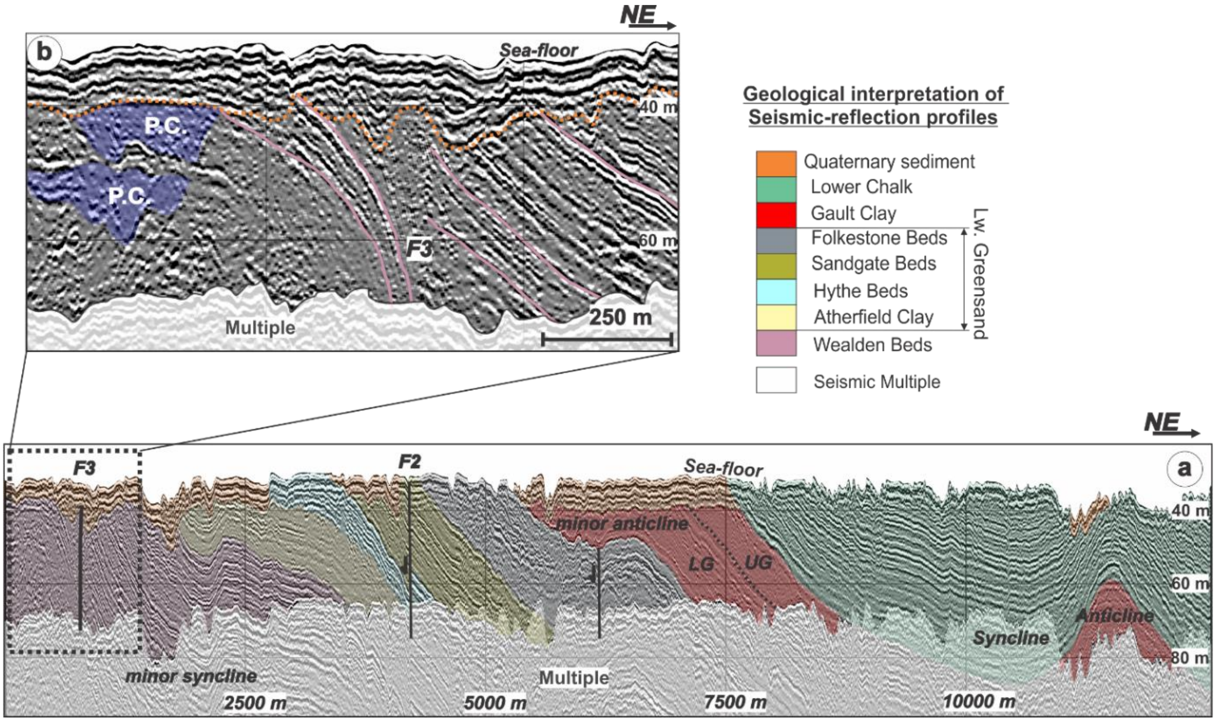

Figure 4.7. (a) Interpreted single-channel seismic-reflection profile (see Figure 4.6 for location). (b) Enlarged view of the rectangle (dash outline) indicated in (a). P.C.: possible Wealden buried palaeo-channels; pale violet lines in (b): seismic horizons selected to illustrate the strata geometry around F3. The depth conversion in this and the following interpreted seismic profiles is based on a mean constant velocity of 2000 m s^{-1}.

(4) Finally, the marine Chalk Group, which is generally subdivided in three units (Hamblin et al., 1992; Mortimore, 2011), i.e. the Lower, Middle and Upper Chalk formations. The Lower Chalk Formation is especially easy to identify in the seismic reflection profiles thanks to the double diffraction produced by the Channel Tunnel (Figure 4.8), which was bored into it (Warren and Harris, 1996). Moreover, the lower boundary of the Lower Chalk Formation is highlighted by the contrast between the parallel reflections of moderate amplitude characterizing its seismic facies and the almost acoustically transparent Upper Gault Clay that lies underneath (Figure 4.7a). The boundaries between the three Chalk formations themselves are much less evident because their seismic facies vary significantly from one profile to another. In this study, we have therefore mapped these boundaries by combining our seismic reflection and bathymetric datasets with the BRGM borehole descriptions and the bedrock-geological map shown in Figure 4.3.

All of the sedimentary formations described above outcrop at the seafloor, where they define several subparallel WNW–ESE ridges and sharp slopes resulting from differential erosion (Figure 4.5, Figure 4.6 and Figure 4.8). Especially marked in the bathymetry are some of the units composing the Lower Greensand and Chalk Groups, which withstand erosion better than the Gault Clay and Wealden Beds formations. For instance, the Hythe Beds form a prominent ridge (in this paper referred to as “Greensand ridge”) on the seafloor that can be followed over almost the entire width of the Dover Strait (see Figure 4.8 and Figure 4.9). The different response to erosion of the Cretaceous sedimentary formations outcropping at the seafloor has also emphasized some old tectonic-related features, like minor folds and fault scarps (see Figure 4.6a and Figure 4.8). In addition to the outcropping bedrock, the seafloor presents two prominent Quaternary erosional/sedimentary features: the aforementioned broad NE–SW-oriented palaeo-channel, known as the Lobourg Channel, and two kilometer-scale, elongated, subparallel sandbanks, called The Varne and The Colbart (Figure 4.4). The latter two are associated with postglacial Holocene sea level rise (Reynaud et al., 2003). In addition to these major sedimentary/erosional features, the bathymetry shows several minor Quaternary sandbanks and erosional features (troughs, grooves, channels, etc.) carved in the seafloor over the entire study area (see Figure 4.5, Figure 4.6 and Figure 4.8).

4.4.2 Tectonic interpretation

The geometry of the Cretaceous strata outcropping in the Dover Strait is strongly controlled by several WNW–ESE-oriented, subparallel minor and major synclines, anticlines and fault systems (Figure 4.5 – Figure 4.12). We can distinguish two deformation styles: (1) a broad asymmetric anticline/syncline system, a north-facing monocline structure and a reverse fault (F1), all three accommodating apparent compressional deformation; and (2) several faults (i.e. F2–8) accommodating vertical offsets more typical of extensional or strike-slip settings.

The north-facing monocline structure traverses the almost-entire width of the Dover Strait with its axis coinciding on the seafloor with the WNW-striking Greensand ridge (Figure 4.8). This structure is without doubt the most significant deformational feature located in the Dover Strait, in which the otherwise sub-horizontal Cretaceous strata are significantly tilted to the east. The monocline structure accommodates most of the deformation documented by the seismic investigation. In fact, the vertical offset produced by this structure is obscured in the seismic reflection profiles by the first seismic multiple. This means that it must be larger than the maximum vertical penetration of the single-channel seismic-reflection data. We can thus assume offsets greater than 100 m (see Figure 4.8).

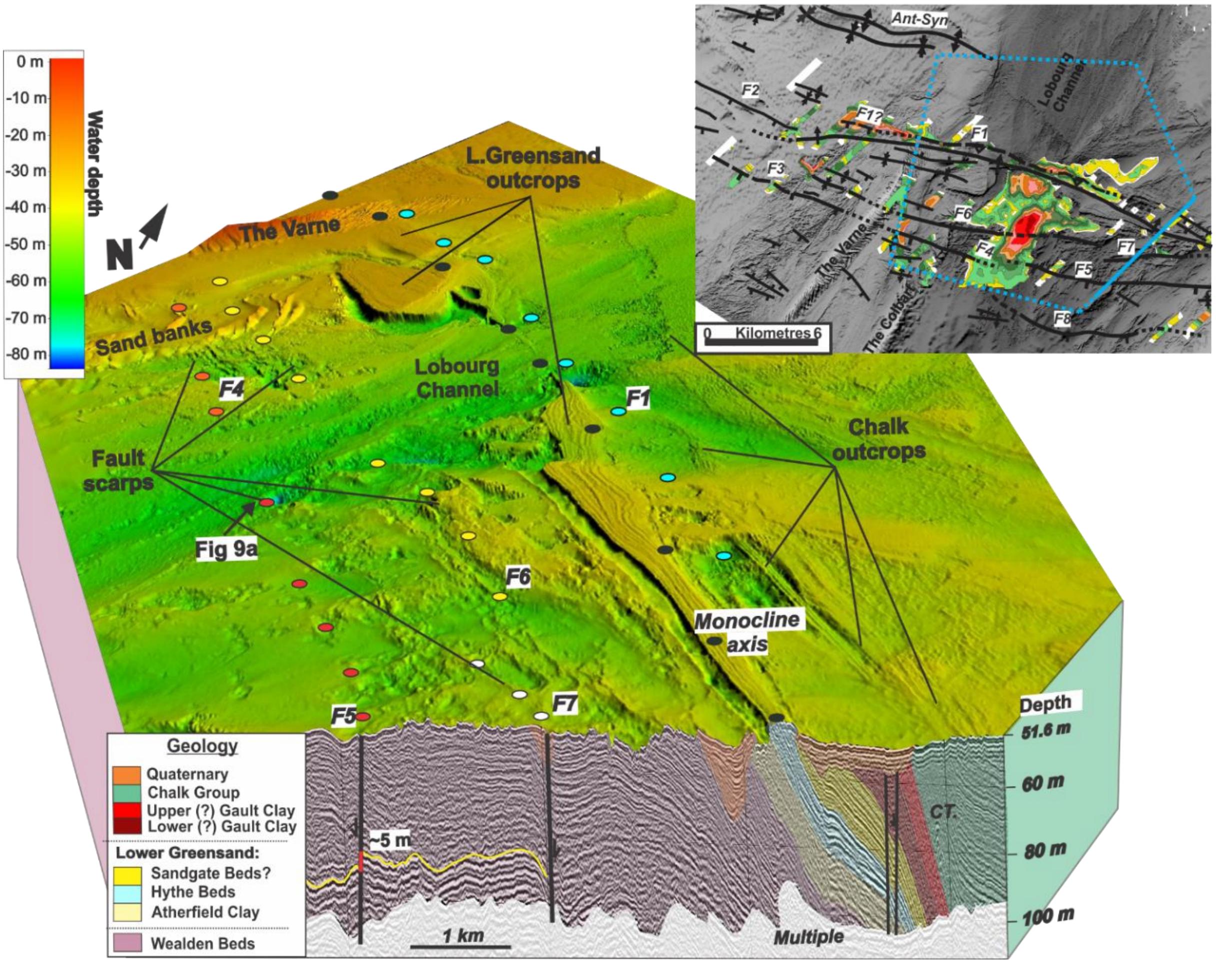

Figure 4.8. 3-D block showing the central and southeastern merged bathymetry in relationship with the interpretation of one of the single-channel seismic-reflection profiles. Colored circles in the bathymetry represent the fault traces and monocline axis inferred from the seismic investigation. CT: hyperbolic diffraction due to the Channel tunnel. Inset: structural map with the area shown in the 3-D block (dash line) indicated. Continuous blue line: location of the seismic reflection profile. See Figure 4.6 for the Inset depth color scale, shade scale of isopach map and structural interpretation.

F1 is located in the central part of the strait, running parallel to the monocline structure over more than 13 km (Figure 4.5 and Figure 4.6). This high-angle (60–80◦) west-dipping fault juxtaposes the Gault Clay and Lower Chalk formations against the lower units of the Lower Greensand (Figure 4.9b and Figure 4.9c), suggesting significant reverse offsets. F1 rarely reaches the seafloor due to the presence of a buried, 70–80 m deep, palaeo-depression carved along its almost-entire length (Figure 4.9). The presence of this palaeo-depression and the high position of the seismic multiple prevented the direct measurement of the vertical offset induced in the Cretaceous bedrock by this structure. A rough estimation of 40–50 m can be calculated though assuming that the thickness of the Lower Greensand Formation does not vary too much between the seismic reflection profiles shown in Figure 4.8 and Figure 4.9(c) (distance ~5 km).

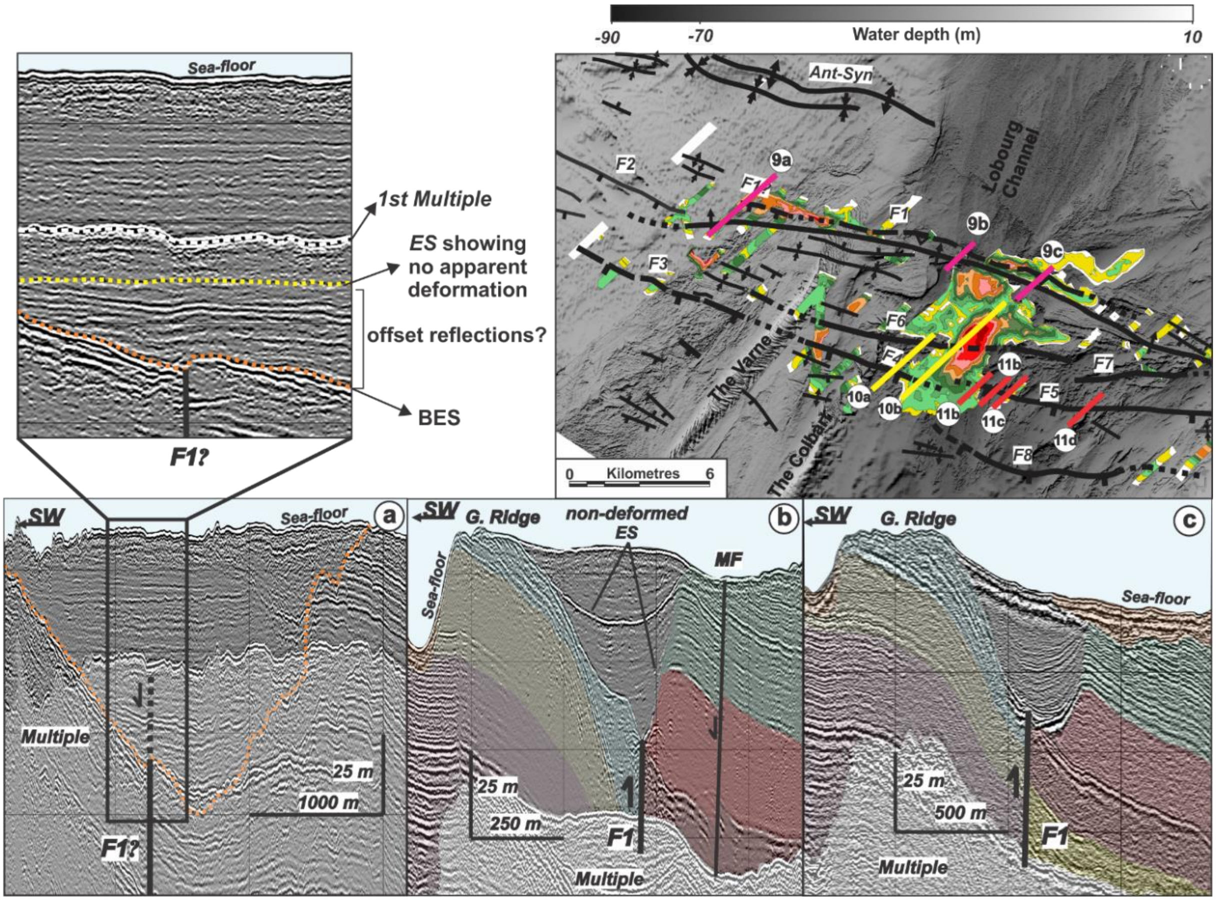

Figure 4.9. Selected parts of three single-channel seismic-reflection profiles (dark pink lines in structural map) acquired across fault F1. G. Ridge: Greensand ridge; BES: basal erosional surface; ES: erosional surface; MF: minor fault (only identified in this profile). See Figure 4.6 and Figure 4.7 for shade scale of isopach map and geological interpretation of the seismic reflection profiles, respectively. Yellow and red lines in the structural map indicate the location of the seismic reflection profiles shown in Figure 4.10 and in Figure 4.11 (a)–(d), respectively.

F1 seems to terminate to the northwest at one of the Fosses Dangeard palaeo-depressions (Figure 4.9a). At that location, the BES and lower internal strata of this palaeo-depression are ~5 m offset to the southwest (Figure 4.9a). This suggests normal or strike-slip faulting instead of the reverse deformation observed along F1 farther to the southeast (Figure 4.9c). This might indicate, on the one hand, that F1 accommodated some normal or/and strike-slip deformation following the formation of the Fosses Dangeard. On the other hand, it is also possible that we are looking at two different faults presenting opposite deformational style. In any case, the offset disappears above the first erosional surface that truncates the lower layers of the palaeo-depression infill (Figure 4.9a). Other troughs and small-buried scarps are seen farther to the southeast at the intersection of the Quaternary palaeo-depression and F1 (Figure 4.9b and Figure 4.9c). These features are also limited to the BES and lowermost strata infilling the palaeo-depression and do not show any predominant deformation style.

The anticline/syncline system is located ∼6 km to the northeast of the monocline/F1 structure presenting a similar trend and lateral extent as F1 (see Figure 4.5 – Figure 4.7). Besides some minor non-deformed Quaternary palaeo-channels carved into this structure, deposits younger than Upper Cretaceous have been eroded or not deposited on top of it, preventing the assessment of the complete activity history of this feature.

The second set of tectonic structures, that is those accommodating apparent extensional and/or strike-slip deformations, is composed of several minor and major high-angle (60–90◦) fault systems. The most significant are: the WNW–ESE trending faults F3 and F4 and the W–E trending faults F5, F6, F7 and F8 (see Figure 4.5 to Figure 4.12). Faults F3, F4 and F5 show similar geometries and characteristics, which suggests that they are probably different segments of the same system. All three faults dip to the southwest and offset normally the Wealden Beds. The offsets produced by these faults vary both laterally and with depth, ranging between 5 and 40 m. This is well evidenced in Figure 4.11, where we observed an increase of the offset from northwest to southeast from ∼9 to ∼24 m in less than 2 km along strike. The offset decreases again to ∼5 m 3.5 km further to the southeast (Figure 4.11d). We have also noticed that the Wealden strata located to the southwest (hanging wall) of F5 is generally downwarped near the fault plane. This downward bending of the strata varies significantly along strike and with depth, being sometimes more pronounced in the upper seismic units than in the lower ones (Figure 4.11b).

The fault system F3–F4–F5 presents minor scarps in the outcropping Wealden Beds (Figure 4.6a). However, it does not seem to reach the seabed surface through the Quaternary sediment-filled palaeo-depressions and palaeo-channels. Rather, the possible Quaternary deformations accommodated by these faults are limited to the BES and lowermost infill of the palaeo-depressions (see Figure 4.7b and Figure 4.10b). Only F5 seems to extend locally all the way to the seafloor along the western slope of a partially filled depression carved into the Lobourg Channel (Figure 4.8 and Figure 4.11a). Nevertheless, despite the clear throws observed at depth in seismic reflection profiles acquired less than 200 m apart, the scarp associated with F5 is restricted to the partially filled depression. No other scarp or offset have been observed within the Lobourg Channel or any of the other sediment-filled palaeo-depressions traversed by this fault system (see Figure 4.7b, Figure 4.10b and Figure 4.11). In fact, deformations possibly linked to the tectonic activity of the system F3–F4–F5 at other locations consists of small changes in the geometry of the palaeo-depression's BES and lower infill near to or across the fault planes (e.g. Figure 4.7b). This suggests that F5 was exhumed at that location by a concentration of the scouring rather than by fault activity.

Faults F6 and F7 are both high angle E–W trending faults separated by a right step-over of 1.4 km (Figure 4.5). These faults present localized minor scarps on the seafloor within outcropping Wealden Beds and in locations with thin Quaternary sedimentary cover (Figure 4.8). The scarps are not continuous through the Lobourg Channel, where the erosional/depositional features located in it appear unaffected by these faults (Figure 4.8). F6 seems to be related to a minor step or buried scarp disrupting the southwestern slope of one of the palaeo-depressions located underneath the Lobourg channel (Figure 4.10b). However, the low amplitude of the reflections infilling this palaeo-depression prevented the characterization of the possible deformation. In any case, it does not seem to extend beyond the thin package of high amplitude reflections corresponding to an erosional surface carved into the sedimentary infill (see Figure 4.10b).

The sense and amount of vertical deformation induced by faults F6 and F7 in the Cretaceous bedrock is also uncertain. We were unable to estimate either of these parameters due to the seismic multiple and the strong differences between the seismic facies at either side of the fault planes (Figure 4.8 and Figure 4.10a). It is actually possible that this fault system has accommodated significant horizontal offset, which would explain our inability to correlate the seismic reflections across the fault. Indeed, the Greensand ridge (axis of the monocline) appears to have been right-laterally bent about 350 m where it intersects the inferred continuation of F7 (Figure 4.5). Unfortunately, we did not find any other right-laterally deformed bathymetric markers to positively link this bend with activity on the fault.

Finally, F8 is only imaged in one of the single-channel seismic-reflection profiles (Figure 4.12) and two of the multichannel seismic lines, preventing an accurate analysis of its activity. This nearly W–E trending southwest-dipping fault appears to offset normally the lower Cretaceous and probably upper Jurassic strata by about 20 m (Figure 4.12). No Quaternary deformation has been observed in the available geophysical data associated with this structure.

Apart from the tectonic structures described above, several other faults and folds are present in our study area deforming the Cretaceous bedrock (see Figure 4.5 to Figure 4.11). We do not describe these structures in detail because they are either of minor importance (e.g. F2) with respect to the ones described above or poorly imaged in the seismic investigation (e.g. F3b). However, these structures are included in the structural and geological maps shown in Figure 4.5 and Figure 4.12.

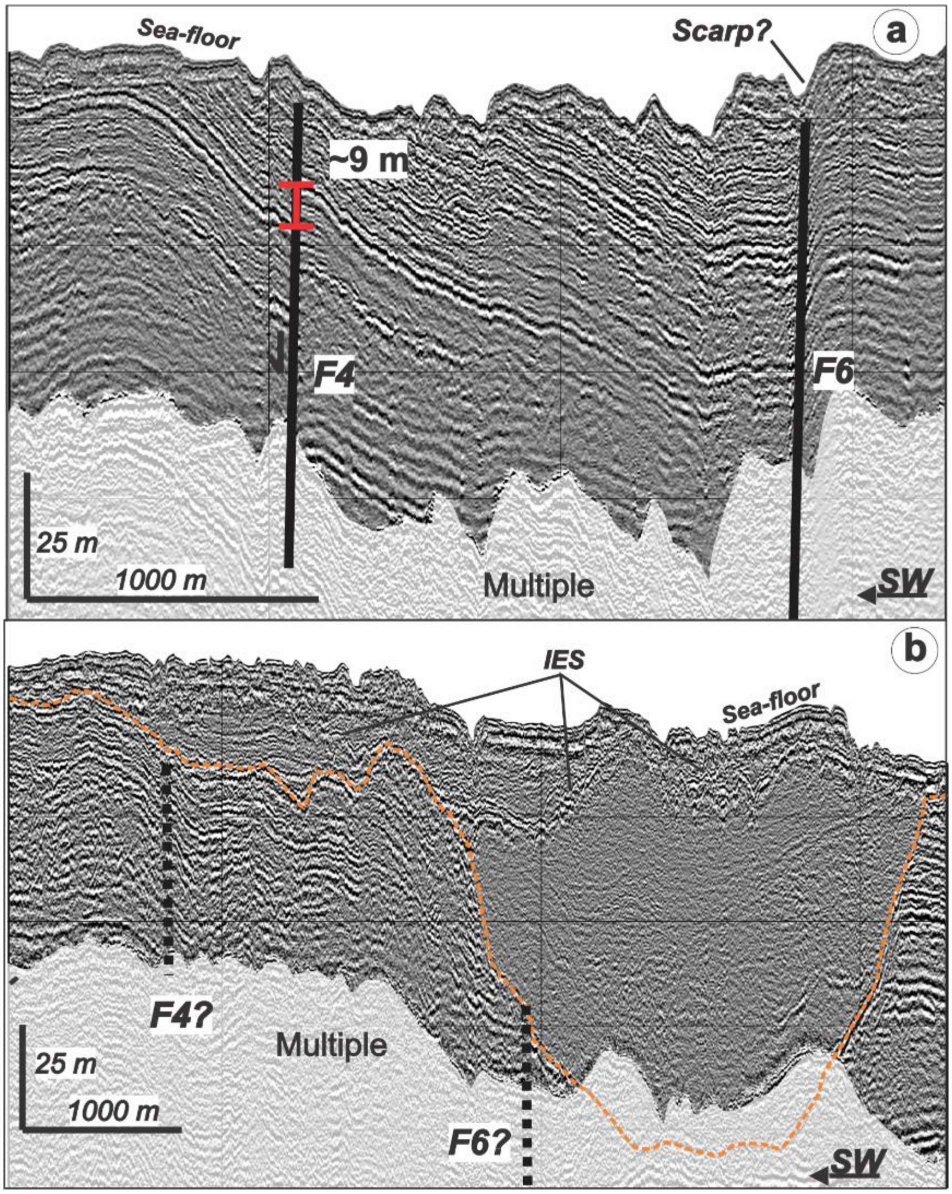

Figure 4.10. Parts of two single-channel seismic-reflection profiles (location indicated in Figure 4.9) traversing faults F4 and F6, and a major palaeo-depression of the Fosses Dangeard. Orange dashed line in (b): palaeo-depression BES.

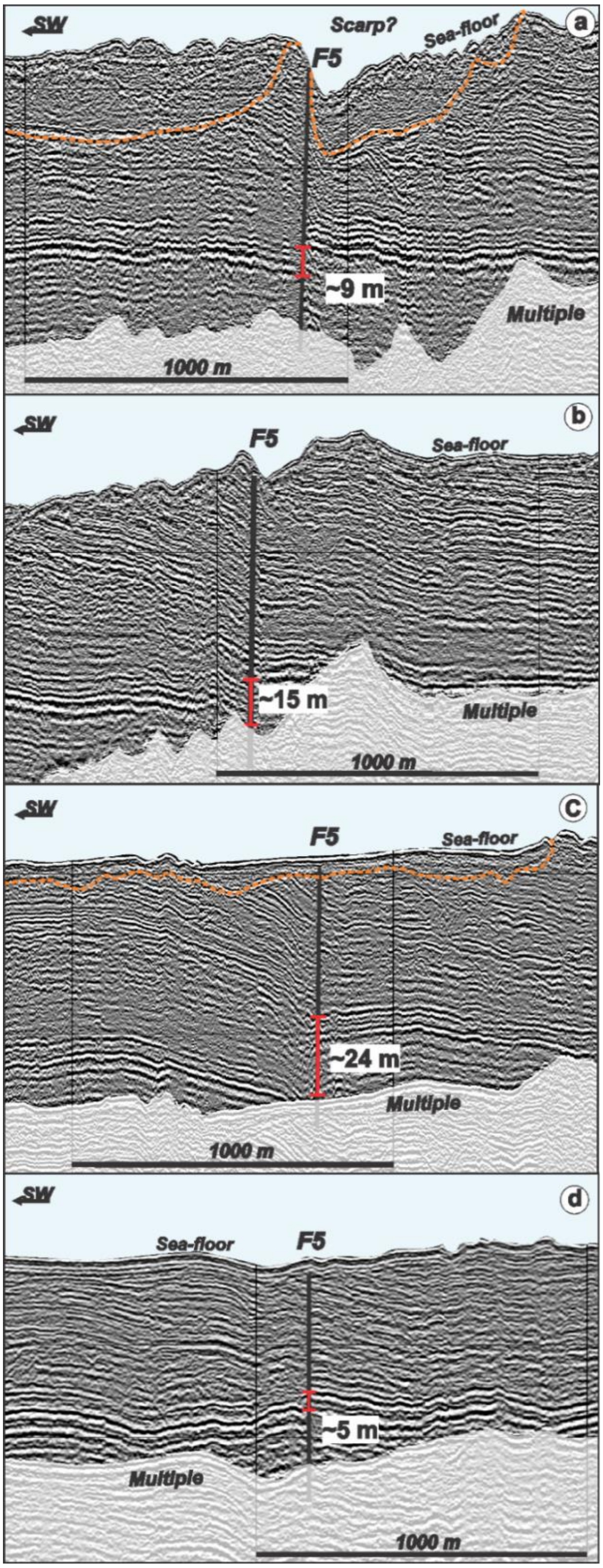

Figure 4.11. Selected parts of four single-channel seismic-reflection profiles acquired across F5 (location indicated in Figure 4.9). Red lines in (a), (b), (c) and (d): offsets induced in the Wealden strata by F5; orange dash line: Quaternary palaeo-depressions BES.

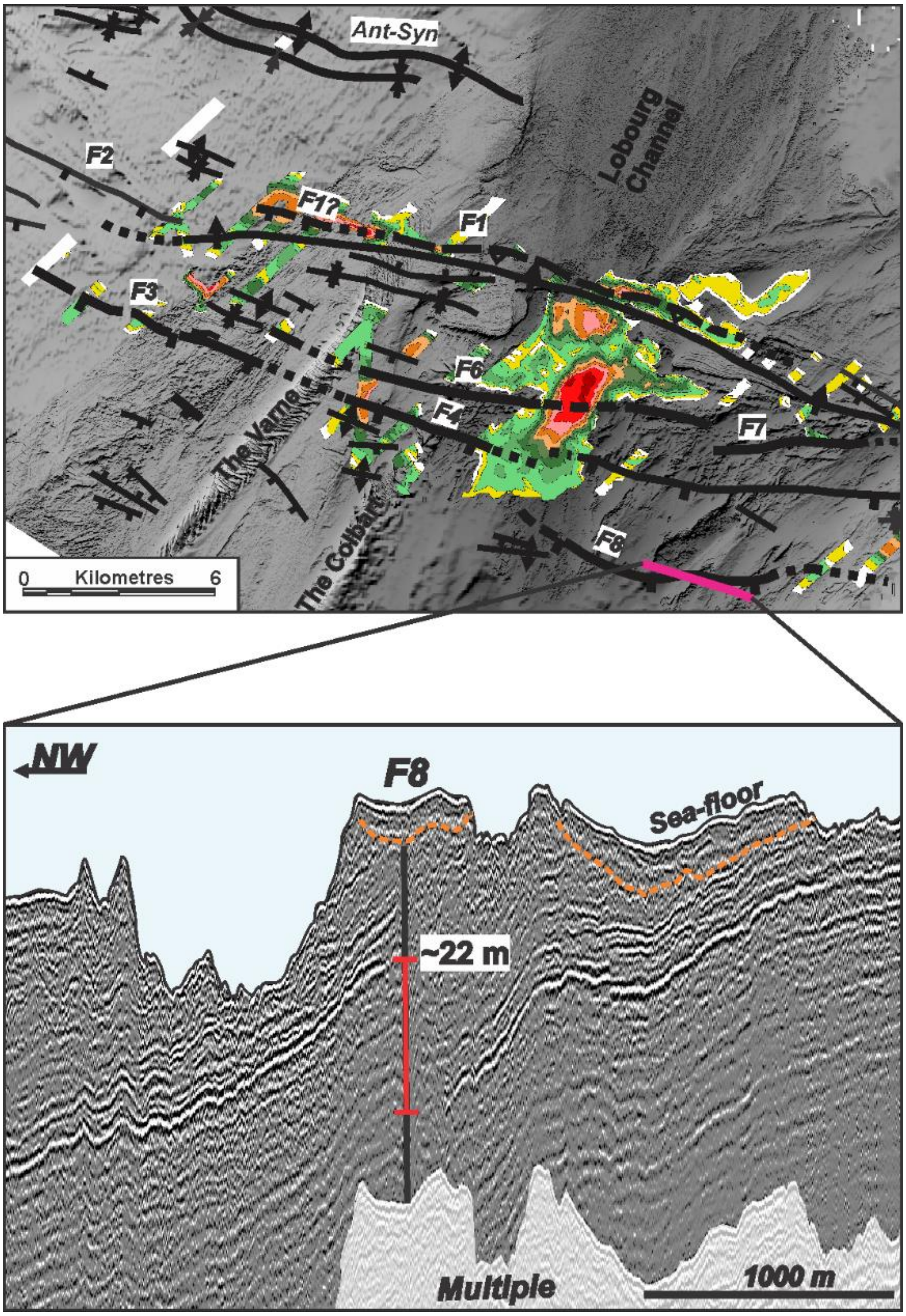

Figure 4.12. Selected parts of the single-channel seismic-reflection profile acquired across F8. Red line indicates the offset induced in the Wealden strata by F8; orange dash line: possible BES of Quaternary deposits.

4.5 Discussion

The new bathymetric data gathered for this study have revealed a seafloor geomorphology strongly influenced by the erosional and sediment-starved conditions currently prevailing in the Dover Strait. Indeed, the geophysical data shows that, apart from some minor and major sandbanks, there has been almost no Holocene deposition in this area. This environment has resulted in very good preservation of the major Late-Quaternary erosional features (e.g. the Lobourg Channel) and the exposure of the bedrock geology and tectonic structures in the seafloor (see Figure 4.5 – Figure 4.12).

The Quaternary sediment-filled Fosses Dangeard are located along or in between the main tectonic structures identified in this area. Importantly, several of the deepest buried depressions were carved into non-deformed, horizontally stratified bedrock. Therefore, even though some of the Quaternary palaeo-depressions present possible minor deformations affecting their BES, neither the incision of the Fosses Dangeard nor their morphologies were induced by tectonic forcing or significantly controlled by the tectonic structures traversing the Dover Strait (Figure 4.6, Figure 4.9 and Figure 4.10). These features were thus formed by erosion as already stated by most of the studies previously performed in this area (e.g. Destombes et al., 1975; Smith, 1985; Hamblin et al., 1992; Gupta et al., 2007).

The tectonic structures identified in the Dover Strait are mainly deforming Lower Cretaceous sedimentary formations. The largest deformations are concentrated on the WNW–ESE north-facing monocline and fault F1. These two structures present the same apparent polarity (i.e. F1 reverse offset is in agreement with the monocline folding), suggesting that their associated deformations were produced during a similar tectonic regime. Their proximity and the continuity of the Greensand ridge (which is the surface expression of both the monocline and hanging wall of F1) indicate that these two deformational features are probably different expressions of the same structure. In other words, the monocline is most likely a blind reverse fault system, whose master fault or one of its associated segments may reach the surface in the central part of the strait as F1. The monocline/F1 structure, together with the anticline/syncline system identified to its northeast, thus point to a compressional tectonic episode during which the entire Cretaceous sequence was folded and faulted. This episode most likely corresponds to the tectonic inversion that started in early Paleogene times (e.g.

Mansy et al., 2003).

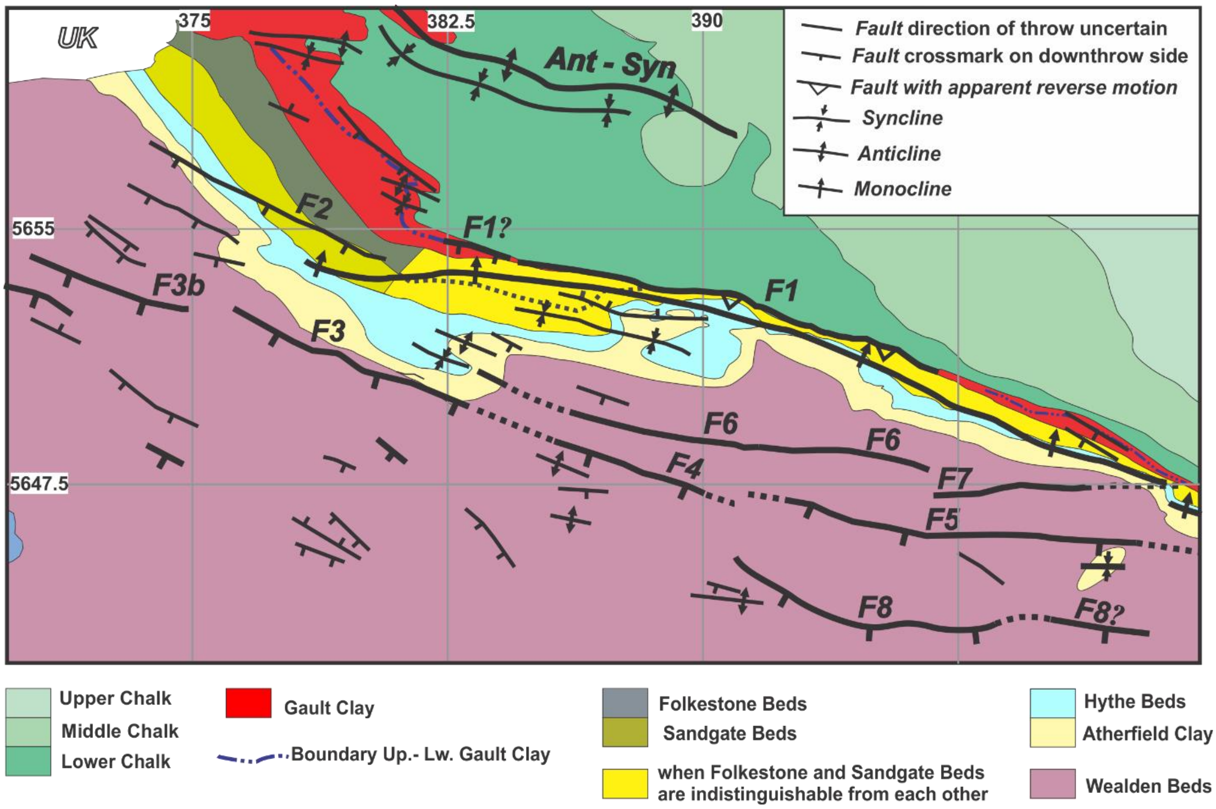

Figure 4.13. Bedrock geology of the Dover Strait derived from the interpretation of the seismic reflection and multibeam bathymetric datasets available for this study.

The offset produced by the monocline/F1 structure in the Cretaceous strata has almost no expression in the seafloor (Figure 4.5), indicating that more than 100 m of bedrock have been eroded from this area since these structures were formed. This has resulted in the exhumation of the Wealden Beds to the south of the north-facing monocline (Figure 4.8 and Figure 4.9). Younger Cretaceous formations outcrop only northwestward of this structure. Notably, the Lower Greensand is restricted to the monocline axis in the central part of the Strait. This observation disagrees with the bedrock-geological map currently available, which proposed a broader outcrop for the Lower Greensand Formation (see Figure 4.3). Neither does that bedrock-geological map indicate the significant reverse offset produced by F1 in the central part of the strait. Figure 4.13 shows the new bedrock-geological map derived from our investigation, including the aforementioned observations and the lithology and thickness variations of the Lower Greensand and Gault Clay formations described in Section 4.1.

The compressional deformation produced by the monocline/F1 structure does not seem to have influenced the shape of the Quaternary palaeo-depressions located in this area. On the contrary, the Quaternary offset associated with the possible northwestern continuation of F1 follows the sense of its dip, resulting in geometries more typical of extensional faults (Figure 4.9a). That suggests that the compressional regime ended prior to the formation of the Fosses Dangeard.

Significant deformation of the Lower Cretaceous strata has also been observed along fault systems F3–F4–F5, F6–F7 and F8, most of them associated with apparent normal and/or strike-slip faulting. A characteristic of these faults is the significant lateral variation of their associated offsets and the vertical and horizontal variations of the bending of the strata near their fault planes (see Figure 4.7, Figure 4.10 and Figure 4.11). Similar variations of the offset and seismic-strata geometries have been observed in the central English Channel by Collier et al. (2006). These authors interpreted that as the consequence of the tectonic inversion of initially syn-sedimentary extensional faults combined with the heterogeneous sediment composition and spatial distribution of the Lower Cretaceous continental deposits. This interpretation is consistent with the lateral and vertical variation of the offsets and bending of the strata observed along the system F3–F4–F5. The fact that we do not observe an inversion of the original normal offsets associated with these structures is because they only traverse Wealden and older sedimentary formations in our study area. That is, the reverse deformation produced in the Wealden Beds during the Tertiary tectonic inversion may not have been enough to compensate the significant pre-existing extensional offsets. On the other hand, the monocline/F1 and anticline/syncline systems are deforming much younger sediments (Aptian–Turonian age), which were subjected to Cretaceous extensional settings during shorter periods of time than the Wealden Beds. Hence, their original normal offsets would have been much smaller and, so, overcompensated by the Tertiary tectonic inversion. This is consistent with the observations of Minguely et al. (2010) and Underhill and Paterson (1998) in the North Artois Shear Zone (northeastern France) and the Wessex Basin (Southern England), respectively. Both studies showed several examples of faults that offset the upper and younger sedimentary strata reversely, while the offset in the lower and older strata remains normal.

Alternatively, the particular characteristics of the fault systems F3–F4–F5 and F6–F7 and the apparent change of polarity along F1 could be due to strike-slip deformation. This interpretation would be in good agreement with the high dipping angles of F6 and F7 fault planes and the apparent right-lateral bend of the Greensand ridge observed at its intersection with the possible prolongation of F7 (Figure 4.5). The right-lateral offset of the monocline structure would actually imply that this possible strike-slip motion took place following the Tertiary compressional episode. If that was so, it is plausible that the system F3–F4–F5 may also have accommodated some strike-slip deformation following the tectonic inversion. However, the observed up-sequence variations in dip of the seismic-strata near the system F3–F4–F5 fault planes are better explained by the tectonic–inversion hypothesis.

The limited penetration of the seismic reflection profiles available for this study prevented imaging the basement and the relationship among the different faults at depth. It is remarkable though that all major structures are located within the gravity anomaly interpreted as the Sangatte Fault (compare Figure 4.2 and Figure 4.5), suggesting that they all belong to the same structure: the Sangatte Fault system.

We have not been able to identify the actual source of the AD 1580 or the possible deformation associated with the rupture that caused this earthquake. However, we have identified possible Quaternary fault activity from irregularities, offsets and changes of the dip within the lower sedimentary infill and BES of Pleistocene palaeo-depressions (Figure 4.7 to Figure 4.11). Especially notable are the irregular cross-sectional morphologies of some minor palaeo-depressions located along F3 (Figure 4.7b) and the apparent ∼5 m vertical offset of one of the palaeo-depression BES by the possible northwestern continuation of F1 (Figure 4.9a). Nevertheless, neither of these deformations seems to extend above the first internal erosional surfaces that truncates the lower layers of the infills of the Fosses Dangeard (see Figure 4.7b, Figure 4.9 and Figure 4.10b). In addition, these faults have virtually no expression on the seafloor where their paths cross the various Quaternary palaeo-depressions and palaeo-channels. This means that the deformation cumulated by these faults since the incision of the first internal erosional surface identified in the infill of the Fosses Dangeard must be below the maximum resolution of the geophysical data, which is 1–3 m for the seismic reflection profiles and 1.5 m for the bathymetric data.

The Dover Strait was exposed to subaerial erosional/depositional settings during the last three major Pleistocene glacial periods (e.g. Gibbard, 1995). Consequently, we cannot rule out the possibility that the erosional surfaces that are unaffected by the faults were formed during the last glacial period (0.11–0.012 Ma). This is also true for the apparent lack of deformation in the seafloor. Indeed, any minor deformations could have been washed out during the last marine lowstand; especially those affecting the Lobourg Channel, which may have confined a significant river during the Last Glacial Maximum (e.g. Gibbard, 1995). Therefore, with the data available for this study we can only state that the Sangatte Fault system has produced deformations of less than 1–3 m since the Holocene started. The cumulated offset since the Fosses Dangeard were formed (possibly 0.45 Ma BP) appears to be of ~5m. However, this could be smaller, as the buried scarp on which we measured this offset could have been caused or exaggerated by the erosional process that originated the Fosses Dangeard.

In summary, earthquakes of magnitudes equal to or higher than 6 seem to have been very rare in this area since Middle Pleistocene. This makes the AD 1580 an exceptional event. In fact, assuming that the cumulated deformation accommodated by the Sangatte Fault system during the Holocene epoch is just below the maximal resolution of the bathymetric data (1.5 m), we can estimate by applying the equations of Wells and Coppersmith (1994) that this deformation would correspond either to the occurrence of a single Mw ~7.0 earthquake produced by a fault with an average length of ~40 km, or to very few earthquakes with magnitudes equal to or lower than 6.0 on shorter faults. More importantly, these estimations suggest that none of the earthquakes produced by the Sangatte Fault system since that epoch has exceeded a maximum magnitude of 7.0, which is consistent with the length of the individual structures composing it.

4.6 Conclusions

This study demonstrates the presence of a more complex yet continuous fault-and-fold system traversing the Dover Strait/Pas-de-Calais than previously published. This fault-and-fold system offsets the Cretaceous sedimentary strata enough to influence the bedrock geology outcropping at the seabed. The system is composed of several major and minor tectonic structures accommodating either compressional deformation or deformations more typical of extensional and/or strike-slip regimes. The comparison between these different styles of deformation and the known regional tectonic history suggests that the compressional structures were formed during the Tertiary compressional phase. Apparent normal deformations would be then more likely linked to the Jurassic–Cretaceous extensional phase and/or to a strike-slip episode that took place after the Tertiary tectonic inversion. The latter seems to be the case for the system F6–F7, which may significantly offset right-laterally the compressional monocline structure. However, we have not been able to determine when this possible strike-slip deformation started or whether it has been ongoing during the Quaternary.

The possible Quaternary tectonic activity of the main structures identified in this study seems to have been very low in recent times. Whilst the lack of recent deformation in the bathymetry could be explained by the present strong erosional tidal conditions in the Dover Strait, the fact that we do not see it in the upper layers of the sedimentary infill of the Fosses Dangeard suggests that, if this exists, it must be below the resolution (1–3 m) of the single-channel seismic-reflection data. The cumulated deformation accommodated since Middle Pleistocene by this fault system may actually be lower than 5 m. Therefore, if the Sangatte Fault system has been active during the Quaternary period, the recurrence interval of large earthquakes must have been very long. Thus, earthquakes similar to the AD 1580 event have been rare in this area. Earthquakes of greater magnitudes seem to have been even rarer or non-existent. Nonetheless, the size of the structures identified in this study suggests that they could potentially rupture in earthquakes with magnitudes up to 7.

In conclusion, this study supports the episodic character of the seismic activity typical of tectonic structures located in plate interiors. These structures present 'short' periods of activity followed by long periods of seismic quiescence that may correspond with the activity of nearby regional structures. This 'migration' of the seismicity is well observed in northwestern Europe, where most of the known moderate and large historical earthquakes have occurred at different locations (Camelbeeck et al. 2007).

References

Arthur, J.C.R., Phillips, G., and McCormick, C.R., 1997, High definition seismic for Channel Tunnel marine route, Modern Geophysics in Engineering Geology. Edited by D. M. McCann, M. Eddleston, P. J. Fenning and G. M. Reeves. Geological Society Engineering Geology, Special Publication, v. 12, p. 327–334.

Bergerat, F., and Vandycke, S., 1994, Paleostress Analysis and Geodynamical Implications of Cretaceous-Tertiary Faulting in Kent and the Boulonnais, Journal of the Geological Society, v. 151, p. 439–448.

Camelbeeck, T., Vanneste, K., Alexandre, P., Verbeeck, K., Petermans, T., Rosset, P., Everaerts, M., Warnant, R., and Van Camp, M., 2007, Relevance of active faulting and seismicity studies to assessments of long-term earthquake activity and maximum magnitude in intraplate northwest Europe, between the Lower Rhine Embayment and the North Sea, The Geological Society of

America, v. 425, p. 193–224.

Chadwick, R.A., Evans, D.J. and Holliday, D.W., 1993, The Maryport fault: the post-Caledonian tectonic history of southern Britain in microcosm, Journal of the Geological Society, v. 150, p. 247 – 250.

Colbeaux, J.P., Somme, J., and Tuffreau, 1981. Tectonique quaternaire dans le Nord de la France: l'apport du gisement paléolithique de Biache-Saint-Vaast, Bulletin de l'Association française pour l'étude du quaternaire, v. 18, p. 183–192.

Collier, J.S, Gupta, S., Potter, G., and Palmer-Felgate, A., 2006, Using bathymetry to identify basin inversion structures on the English Channel shelf, Geology, v. 34, p. 1001–1004.

Collier, J. S., Oggioni, F., Gupta, S., García-Moreno, D., Trentesaux, A., De Batist, M., 2015, Streamlined islands and the English Channel megaflood hypothesis, Global and Planetary Change, v. 135, p. 190–206.

De Béthune, P., and Bouckaert, L., 1968, Geological Map of Belgium and Neighbouring Countries: Brussels, Geological survey of Belgium Miscellaneous Geological Maps, scale 1:2,000,000, 1 sheet.

Destombes, J. P., Shephard-Thorn, E. R., Redding, J. H., and Morzadeckerfourn, M. T., 1975, Buried valley system in the Strait of Dover, Philosophical Transactions of the Royal Society A-Mathematical Physical and Engineering Sciences, v. 279, p. 243–256.

Ehlers, J., and Gibbard, P. L. (Eds.), 2004, Quaternary glaciations-extent and chronology: part I: Europe. Developments in Quaternary Science, v. 2, Elsevier, Amsterdam.

Everaerts, M., and Mansy, J.-L., 2001, Le filtrage des anomalies gravimétriques; une clé pour la compréhension des structures tectoniques du Boulonnais et de l'Artois (France), Bulletin de la Société Géologique de France, v. 172, p. 267–274.

Froidevaux, C., Paquin, C., and Souriau, M., 1980, Tectonic stresses in France: in situ Measurements with a flat jack, Journal of Geophysical Research, v. 85, p. 6342–6346.

Gibbard, P.L., 1995. The formation of the Strait of Dover, Geological Society, London, Special Publications, v. 96, p. 15–26.

Gibbard, P.L., and Cohen, K.M., 2015, Quaternary evolution of the North Sea and the English Channel, Proceedings of the Open University Geological Society, v. 1, p. 63–74.

Gibbard, P.L., and Lewin, J., 2016, Filling the North Sea Basin: Cenozoic sediment sources and river styles, Geologica Belgica, v. 19, p. 3–4.

Grünthal, G., Wahlström, R., and Stromeyer, D., 2009, The unified catalogue of earthquakes in central, northern, and northwestern Europe (CENEC) – updated and expanded to the last millennium, Journal of Seismology, v. 13, p. 517–541.

Gupta, S., Collier, J., Palmer-Felgate, A., and Potter, G., 2007, Catastrophic flooding origin of shelf valley systems in the English Channel, Nature, v. 448, p. 342–346.

Hamblin, R.J.O., Crosby, A., Balson, P.S., Jones, S.M., Chadwick, R.A., Penn, I.E., and Athur, M.J., 1992, United Kingdom offshore regional report: the geology of the English Channel, London: HMSO for the British Geological Survey.

Hopson, P.M., Wilkinson, I.P., and Woods, M.A., 2008, A stratigraphical framework for the Lower Cretaceous of England. Research Report RR/08/03, British Geological Survey, Keyworth.

James, J.W.C., Guennoc, P., Harrison, M., Le Bot, S., Philpott, S., Vinchon, C., Bee, E., Simien, F., Janjou, D., Garlan, T., Trentesaux, A., Mahieux, G., Briet D., and Augris C., 2002, GEOSYNTH: A synthesis of the geology and sediments of the Dover Strait and its hinterland. CR/02/078, CDROM, British Geological Survey.

Lericolais, G., Auffret, J.-P., and Bourillet, J.-F., 2003, The Quaternary Channel River: seismic stratigraphy of its palaeo-valleys and deeps, Journal of Quaternary Science, v. 18, p. 245–260.

Mansy J.-L., Manby G.M., Averbuch O., Everaerts M., Bergerat F., Van Vliet-Lanoe B., Lamarche J., and Vandycke S., 2003, Dynamics and inversion of the Mesozoic Basin of the Weald-Boulonnais area: role of basement reactivation, Tectonophysics, 373, 161–179.

Mellett, C., Hodgson, M., Plater, A., Mauz, B., Selby, I., and Lang, A., 2013, Denudation of the continental shelf between Britain and France at the glacial–interglacial timescale, Geomorphology, v. 203, p. 79–96.

Melville, C.P., Levret, A., Alexandre, P., Lambert, J., and Vogt, J., 1996, Historical seismicity of the Strait of Dover – Pas de Calais,Terra Nova, v. 8, p. 626–647.

Minguely, B., Averbuch, O., Patin, M., Rolin, D., Hanot, F., and Bergerat, F., 2010, Inversion tectonics at the northern margin of the Paris basin (northern France): new evidence from seismic profiles and boreholes interpolation in the Artois area, Bulletin de la Société Géologique de France, v. 181, p. 429–442.

Mortimore, R.N., 2011, A Chalk Revolution: what have we done to the Chalk of England? Proceedings of the Geologists' Association, v. 122, p. 232–297.

Musson, R.M.W., 2004, A critical history of British earthquakes, Annals of Geophysics, v. 47, p. 597–609.

Neilson, G., Musson, R.M.W., and Burton, P.W., 1984, The London Earthquake of 1580, April 6, Engineering Geology, v. 20, p. 113–141.

Ottemöller, L., Baptie, B., and Smith, N.J.P., 2009, Source Parameters for the 28 April 2007 Mw 4.0 Earthquake in Folkestone, United Kingdom, Bulletin of the Seismological Society of America, v. 99, p. 1853–1867.

Owen, H.G., 1975, The stratigraphy of the Gault and Upper Greensand of the Weald, Proceedings of the Geologists' Association, v. 86, p. 475–498.

Radley, J.D., and Allen, P., 2012, The Wealden (non-marine Lower Cretaceous) of the Weald Sub-basin, southern England, Proceedings of the Geologists' Association, v. 123, p. 245 – 318.

Reynaud, J-Y., Tessier, B., Auffret, J-P., Berné, S., De Batist, M., Marsset, T., and Walker, P., 2003, The offshore Quaternary sediment bodies of the English Channel and its Western Approaches, Journal of Quaternary Science, v. 18, p. 361–371.

Ruffell, A.H., and Wach, G.D., 1991, Sequence stratigraphic analysis of the Aptian-Albian Lower Greensand in southern England, Marine and Petroleum Geology, v. 8, p. 341–353.

Smith, A.J., 1985. A catastrophic origin for the palaeovalley system of the eastern English Channel, Marine Geology, v. 64, p. 65–75.

Stein S., and Mazzotti S., 2007, Continental Intraplate Earthquakes: Science, Hazard and Policy Issues, The Geological Society of America, Special paper, v. 425, 402 p.

Toucanne, S., Zaragosi, S., Bourillet, J.F., Gibbard, P.L., Eynaud, F., Giraudeau, J., Turon, J.L., Cremer, M., Cortijo, E., Martinez, P., Rossignol, L., 2009, A 1.2 Ma record of glaciation and fluvial discharge from the West European Atlantic margin, Quaternary Science Reviews, v. 28, p. 2974–2981.

Underhill, J.R., and Paterson, S., 1998, Genesis of tectonic inversion structures: seismic evidence for the development of key structures along the Purbeck–Isle of Wight Disturbance, Journal of the Geological Society, London, v. 155, p. 975–992.

Vandycke, S., Bergerat, F., and Dupuis,C., 1988, Paléo-contraintes à la limite Crétacé-Tertiaire dans le Bassin de Mons (Belgique). Implications cinématiques. Relations avec la zone de Cisaillement Nord-Artois, C. R. Académie de Sciences de Paris, v. 307 (II), p. 303–309

Van Vliet-Lanoë, B., Mansy, J.L., Margerel, J.P., Vidier, J.P., Lamarche, J., and Everaerts, M., 1998, Le Pas de Calais un détroit cénozoïque à ouverture multiple, C. R. Académie de Sciences de Paris, v. 326 (II), p. 729–736.

Van Vliet-Lanoë, B., Laurent, M., Bahain, J.L., Balescu, S., Falguères, C., Field, M., Hallégouët, B., and Keen, D.H., 2000, Middle Pleistocene raised beach anomalies in the English Channel: regional and global stratigraphic implications, Journal of Geodynamics, 29, 15–41.

Van Vliet-Lanoë, B., Vandenberghe, N., Bergerat, F., Henriet, J.-P., Mansy, J.L., Mielliez, F., Lacquement, F., Sintubin, M., and Vandycke, S., 2002a, Cenozoic evolution of the Channel and southern North Sea areas (Western Europe). The neotectonic control on sedimentation and rivers. In: Degryse, P. et al. (Ed.) (2002). Contributions to the geology of Belgium and Northwest Europe: Proceedings of the first Geologica Belgica International Meeting, Leuven, 11-15 September 2002. Aardkundige Mededelingen, v. 12, p. 21–24.

Van Vliet-Lanoë B., Vandenberghe N., Laignel B., Laurent M., Lauriat-Rage A., Louwye S., Mansy J.L., Meilliez F., Mercier D., Hallégouët, B., Lacquement, F., Michel, Y., and Moguedet, G., 2002b, Palaeogeographic evolution of Northwestern Europe during the upper Cenozoic, Geodiversitas, v. 24, p. 511–541.

Van Vliet-Lanoë, B., Mansy, J.L., Henriet, J.-P., Laurent, M., and Vidier, J.P., 2004, Une inversion tectonique cénozoïque par étapes : le Pas-de-Calais. Bulletin Société géologique Français, v. 175(II), p. 175–195.

Warren, S.T., and Harris, C.S., 1996, An interpretation of the structural geology. In Engineering Geology of the Channel Tunnel, C. S. Harris, M. B. Hart, P. M. Varley and C. D. Warren (Editors), Thomas Telford Ltd, p. 421–435.

Wells, D.L., and Coppersmith, K.J., 1994, New empirical relationships among magnitude, rupture length, rupture width, rupture area, and surface displacement, Bulletin of the Seismological Society of America, v. 84, p. 974–1002.

Woods, M.A., Wilkinson, P., and Hopson, P. M., 1995, The stratigraphy of the Gault Formation (Middle and Upper Albian) in the BGS Ariesey Borehole, Bedfordshire, Proceedings of the Geologists' Association, v. 106, p. 271– 280.

Zoback, M.D., and Grollimund, B., 2001, Impact of deglaciation on present-day intraplate seismicity in eastern North America and Western Europe. C. R. Académie de Sciences de Paris/Earth and planetary Sciences, v. 333, p. 23–33.

Chapter 5

Two-stage opening of the Dover Strait and the origin of Island Britain

Chapter 5 – Two-stage opening of the Dover Strait and the origin of island Britain

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Abstract

Late Quaternary separation of Britain from mainland Europe is considered to be a consequence of spillover of a large proglacial lake in the Southern North Sea basin. Lake spillover is inferred to have caused breaching of a rock ridge at the Dover Strait, although this hypothesis remains untested. Here we show that opening of the Strait involved at least two major episodes of erosion. Sub-bottom records reveal a remarkable set of sediment-infilled depressions that are deeply incised into bedrock that we interpret as giant plunge pools. These support a model of initial erosion of the Dover Strait by lake overspill, plunge-pool erosion by waterfalls and subsequent dam breaching. Cross-cutting of these landforms by a prominent bedrock-eroded valley that is characterized by features associated with catastrophic flooding indicates final breaching of the Strait by high-magnitude flows. These events set-up conditions for island Britain during sea-level highstands and caused large-scale re-routing of NW European drainage.

Keywords: Geomorphology; Geophysics; Palaeogeography; Northwestern Europe.

Author contributions: S.G. and J.S.C. conceived, designed and coordinated the study. J.S.C. and F.O. processed bathymetry data. G.P. compiled and processed UK single-beam bathymetry data. D.G-M. processed ROB-RCMG bathymetry data. A.T. and B.V.V.L. coordinated RV Sepia II seismic survey. T.C. and M.D.B. designed and coordinated RV Belgica seismic surveys. K.V. and D.G-M. coordinated and conducted data acquisition and processing. J.C.R.A. provided input on Channel Tunnel surveys. S.G., D.G-M., J.S.C. and F.O. analyzed and interpreted geophysical datasets with contributions from A.T. and M.D.B. All authors discussed the results. S.G. and J.S.C. wrote the paper with important contributions from D.G-M.

5.1 Introduction

The geographic insularity of Britain from continental Europe is a consequence of high interglacial sea levels that led to marine flooding of the shallow shelf areas of the English Channel and North Sea (Preece, 1995). Prior to the opening of the Dover Strait, however, Britain remained connected to Europe, even during sea-level highstands, via a structural ridge that extended from southeast England to northwest France. This ridge, made of chalk, comprised the northern limb of the Weald–Artois anticline, and is postulated to have formed a narrow isthmus separating marine embayments to the north (North Sea) and southwest (English Channel) (Gibbard, 1988; Gibbard, 1995). Breaching of this barrier is a necessary prerequisite to form island Britain.

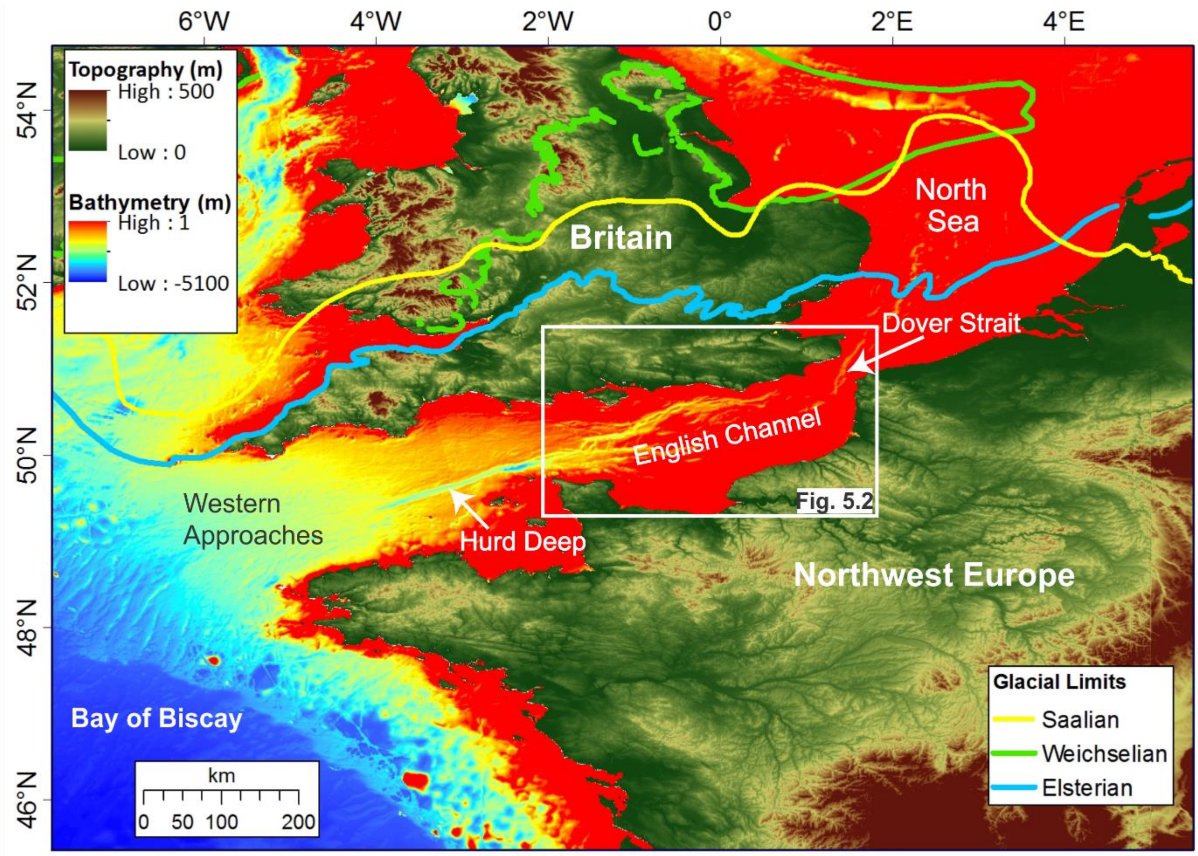


Figure 5.1. Topographic map of the English Channel region plotted with a WGS-1984 projection. The Quaternary glacial limits are from Ehlers et al. (2011). White box indicates areal extent of Figure 5.2. Offshore bathymetry is from GEBCO (www.gebco.net). For this and next figures, onshore elevation is from SRTM (www2.jpl.nasa.gov/srtm/cbanddataproducts.html).

It is widely considered that the breaching of the chalk ridge and opening of the Dover Strait is a consequence of spillover of a proglacial lake that occupied the present-day Southern North Sea basin during the Marine Isotope Stage (MIS) 12 glaciation; i.e. ~450 ka and conventionally equated to the Elsterian–Anglian glaciation (Gibbard, 1995; Smith, 1985; Gupta et al., 2007; Cohen et al., 2014; Collier et al., 2015). Coalescence of the British and Scandinavian ice sheets in the northern and central North Sea basin and the chalk barrier to the south caused fluvial discharge from European rivers and glacial meltwater to become ponded to form this lake (Gibbard, 1988; Cohen et al., 2014). Lake spillover is thought to have released a significant megaflood into a subaerial English Channel (Smith, 1985; Gupta et al., 2007). An extensive network of bedrock-eroded valleys in the central English Channel shows morphologies characteristic of erosion by high magnitude flood flows (Dingwall, 1975; Auffret et al., 1980; Smith, 1985; Lericolais et al., 2003; Collier et al., 2015). Gupta et al. (2007) and Collier et al. (2015) interpreted these valleys as a consequence of catastrophic drainage of a pro-glacial lake. Others contend that while a proglacial lake existed in the southern North Sea, its spillover was not a catastrophic process (Hijma et al., 2012). An alternative model suggests that breaching of the Strait was an incremental process with slow erosion of the Weald-Artois rock barrier at the Dover Strait by fluvial processes during glacial sea-level lowstands and tidal erosion during interglacial highstands (Westaway and Bridgland, 2010; Mellett et al., 2013). In this model, there is no requirement for a proglacial lake in the southern North Sea basin. Thus models for erosion of the palaeovalley networks downstream of the Dover Strait range from those proposing erosion by high-magnitude events (Gupta et al., 2007; Collier et al., 2015) to those suggesting relatively slow fluvial erosion (e.g. Mellett et al., 2013). Testing of models for opening of the Strait has however been limited through lack of high-resolution marine geophysical data at the inferred breach point in the Dover Strait.

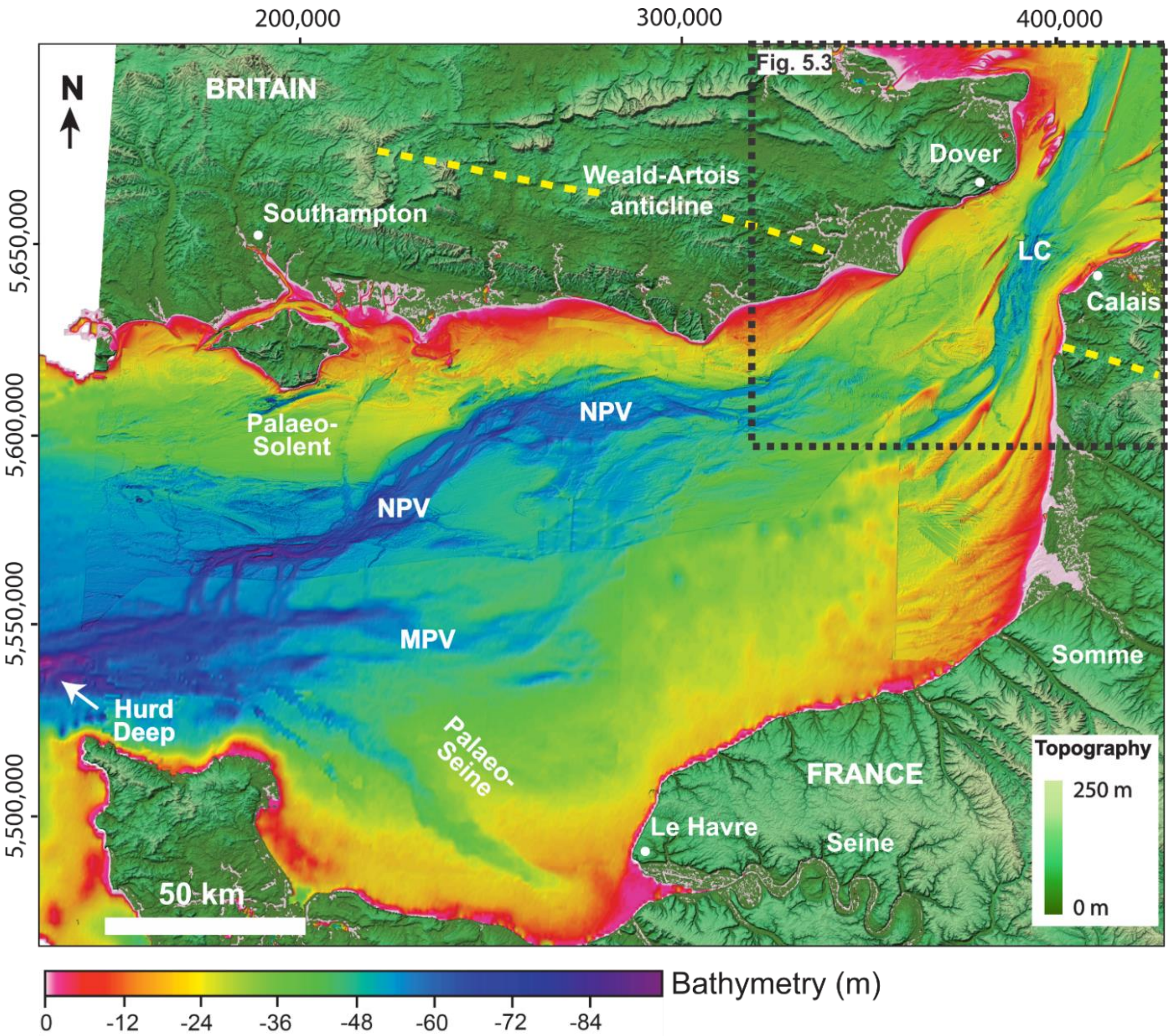


Figure 5.2. Sonar bathymetry of the eastern English Channel shelf gridded at 30 m cell size (see Figure 5.1 for location). The location of the study area in Figure 5.3 and Figure 5.4 is indicated (Black dashed square). LC: Lobourg Channel; MPV: Median palaeovalley; NPV: Northern palaeovalley. An analysis of the downstream morphology is given in Collier et al. (2015). Yellow dashed line indicates the axial trace of the Weald–Artois anticline. Coordinate system for this and the following figures: UTM/N31/WGS-1984/meters.

Early marine geophysical investigations in the center of the Strait in support of the Channel Tunnel engineering project in the 1960s and 1970s identified a set of enigmatic sediment-infilled depressions, termed the Fosses Dangeard (fosse being 'deep' in English) (Destombes et al., 1975; Arthur et al., 1996). At the time of their discovery, glacial erosion by an ice-stream advancing through the Dover Strait was proposed to explain the formation of both the bedrock valley and associated depressions (Destombes et al., 1975; Kellaway et al., 1975). This was however subsequently discounted due to a lack of independent evidence for the presence of ice this far south (Figure 5.1; Ehlers et al., 2011). Smith (1985), in the first proposal of the megaflood hypothesis, speculated that the depressions might represent fossil plunge pools formed at the base of waterfalls overspilling the chalk barrier from a southern North Sea proglacial lake. Erosion of the Lobourg Channel was inferred to represent the final stage in the breaching of the rock dam. The lack of modern high-resolution marine geophysical data in the Dover Strait has meant that the lake overspill and dam breaching model for opening of the Strait is untested.

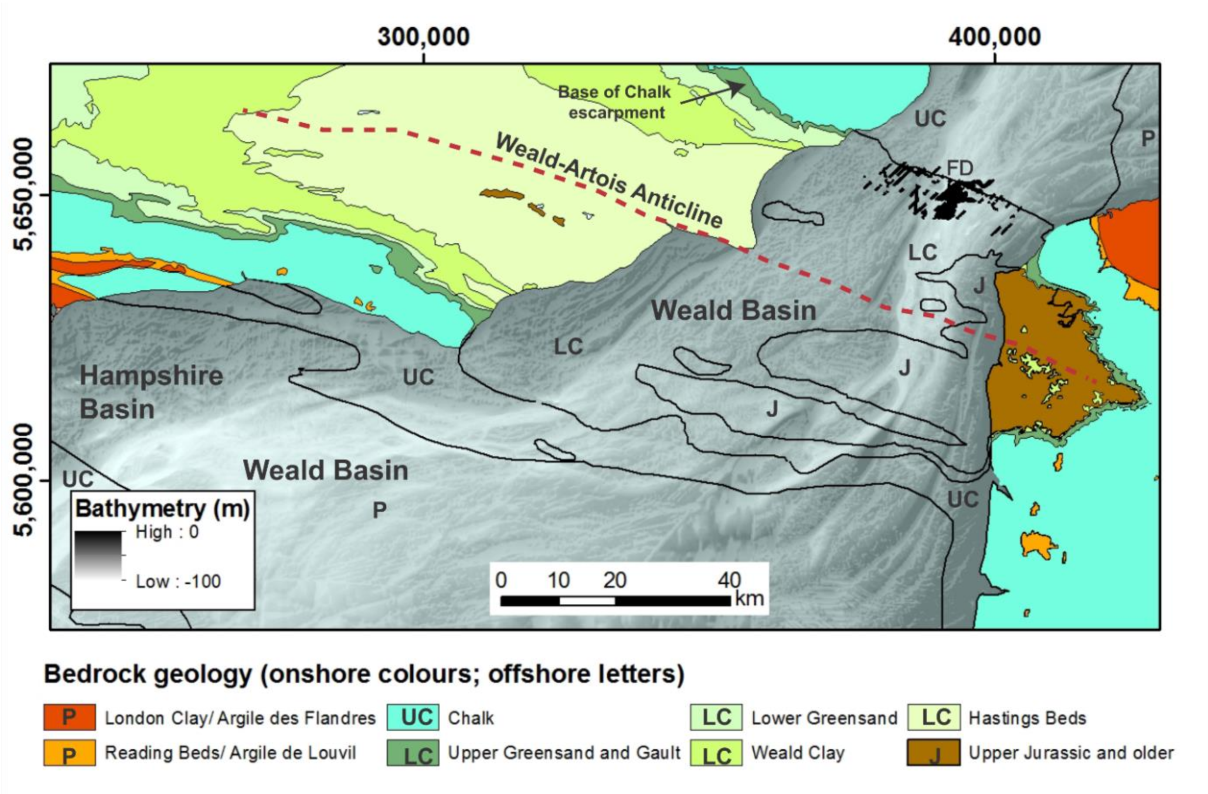


Figure 5.3. Map showing onshore and offshore bedrock geology of Dover Strait area. Onshore bedrock geology shown in colors and offshore geology bedrock geology indicated by letters. The zone of black color in the Dover Strait indicates locations of sediment-infilled Fosses Dangeard along seismic track-lines. Note how the Fosses are localized immediately southwest of the Chalk bedrock outcrop in the northern sector of the Strait. This correlates onshore with the base of the SW-facing escarpment formed by the Chalk (see Figure 5.4). J: Jurassic; LC: Lower Cretaceous; UC: Upper Cretaceous; P: Paleogene.

Here we combine analysis of a regional high-resolution bathymetric grid of the central and eastern English Channel shelf that includes new multibeam sonar data from the Dover Strait, with new high-resolution seismic-reflection data to investigate how the Strait was formed (see Chapters 3 and 4). The mechanism and history of the breaching of the Dover Strait is a question of importance to not only understanding the geographic isolation of Britain from continental Europe (Gibbard, 1995), but also the large-scale re-routing of northwest European drainage, and meltwater to the North Atlantic via the Channel during Pleistocene Glacial periods (Busschers et al., 2007; Busschers et al., 2008; Hijma et al., 2012; Toucanne et al. 2009a; Toucanne et al. 2009b; Toucanne et al. 2010; Plaza-Morlote et al., 2017). Moreover, the opening of the Strait has significance for the biogeography (Preece, 1995; Sutcliffe, 1995; Stuart, 1995; Meijer et al., 1995) and archaeology of NW Europe, with particular attention on the pattern of early human colonization of Britain (Ashton and Lewis, 2002; Ashton and Hosfield, 2010; Ashton et al., 2011; Hijma et al., 2012).

5.2 Methodology

The present investigation combines 2D seismic-reflection data with single-beam and multibeam bathymetric data to interpret the 3D morphology, infill and interrelationship of the Fosses Dangeard and Lobourg Channel. Technical details on the various datasets used in this study are described in Chapter 3 of this dissertation. Location and extension of the various survey grids are shown in Figure 3.3 and Figure 4.4.

The processed seismic data were imported into the interpretation software Kingdom Suite (TKS) and Opendtect (http://www.opendtect.org). The Fosses Dangeard basal erosional surfaces and internal erosional surfaces (when possible) were correlated from one profile to another by using the three-dimensional software Opendtect. Isopach maps were built by subtracting the basal erosional surface of the various infilled depressions from the seafloor and assuming mean seismic velocity through the Fosse infill of 1,800 m s^{-1} (see Arthur et al., 1997). For this calculation, only the infills of the Fosses Dangeard were considered. Major tidal sand ridges (for example, The Varne) were removed by subtracting the seafloor from the erosion surface at the base of the tidal sand ridges. Isopach maps are discontinuous through some of the major depressions (Figure 5.5). This is due to the presence of the first seismic multiples at depths of ~100 m, which obscures the continuation of some of the deepest depressions at depth.

5.3 Results

5.3.1 Seabed geomorphology in the English Channel

A regional-scale compilation of bathymetric data from the central and eastern English Channel (Figure 5.2) shows that a prominent valley, the Lobourg Channel, in the Dover Strait can be traced southwestwards into a network of bedrock-eroded valleys in the central Channel

(James et al., 2002; Gupta et al., 2007; Mellett et al., 2013; Collier et al., 2015). Individual valleys, such as the Northern palaeovalley, have an anabranching planform morphology with channels bifurcating around streamlined bedrock remnants (Gupta et al., 2007; Collier et al., 2015). The continuity of the valley network, in particular the Northern palaeovalley, with the Lobourg Channel in the Dover Strait establishes them as part of the same palaeo-drainage system (Figure 5.4; Collier et al., 2015). For example, the Lobourg Channel at its southwestern extent bifurcates around a prominent bedrock streamlined island (SI 3) to form two channel pathways. Because the valleys in the central Channel are interpreted to have formed by erosion by high-magnitude flood flows (Gupta et al., 2007; Collier et al., 2015), their continuity with the Lobourg Channel suggests that incision of this prominent feature was likely a related process. Below we examine the Fosses Dangeard and the high-resolution morphology of the Lobourg Channel to explore their processes of formation within the context of the English Channel palaeo-drainage system.

5.3.2 Subsurface geomorphology of the Dover Strait.

To investigate the early landscape evolution of the Strait, we used high-resolution seismic profiles to describe the morphology of the Fosses Dangeard in the central part of the Dover Strait. Our lines cover the entire Strait (see Chapters 3 and 4), and so allow us to map the detailed distribution, geometry and internal architecture of the Fosses Dangeard. These data show that the Fosses consist of a series of kilometer–diameter depressions, the base of which are incised up to ∼140 m into Cretaceous bedrock (Figure 5.5a and Figure 5.6). This depth of incision is exceptional given the low gradient setting of the English Channel shelf. Mapping of the Fosses in adjacent seismic lines shows that the deepest parts of the depressions are spatially discontinuous though connected by shallow corridors. In total, we recognize seven major depressions, which we label A–G (Figure 5.5b). Seismic facies in the Fosses infill vary greatly from one Fosse to the next making correlation of seismic units impossible without groundtruthing.

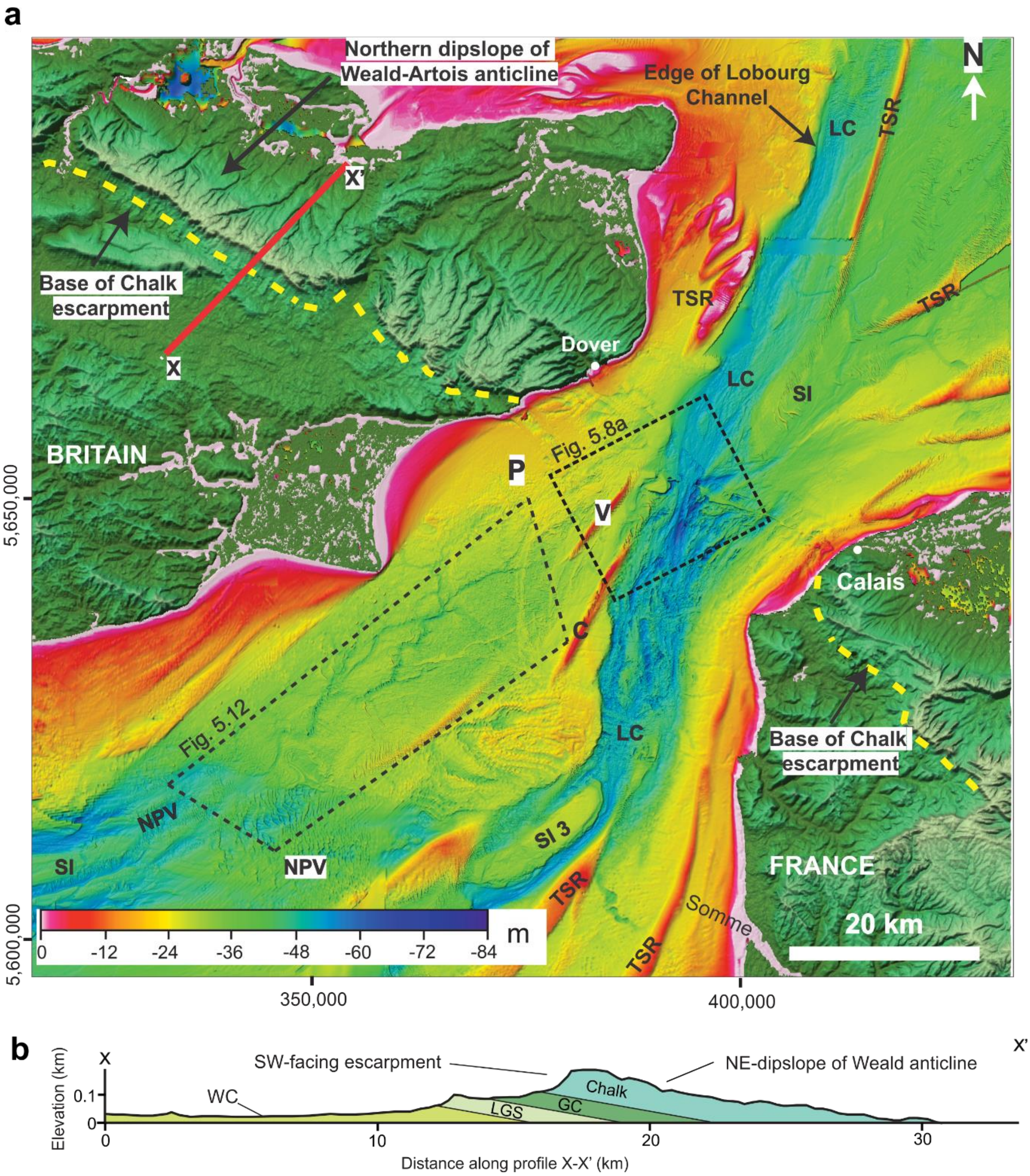


Figure 5.4. (a) Colored and shaded relief bathymetric compilation map of the Dover Strait region. C: Colbart tidal sand ridge; LC: Lobourg Channel; NPV: Northern palaeovalley; P: platform; V: Varne tidal sand ridge; SI: streamlined island; TSR: tidal sand ridge. Note valley network incised into platform, P, indicating sub-aerial exposure of platform surface (see Figure 5.12). Water depth is indicated by color bar. Red line X–X': geological cross-section in b. (b) Geological cross-section across the northern flank of Weald anticline onshore. Note the prominent SW-facing Chalk escarpment and NE-facing gentle dip slope. GC: Gault Clay; LGS: Lower Greensand; WC: Weald Clay.

The cross-sectional geometry of the Fosses Dangeard is illustrated in a seismic reflection profile in Figure 5.5a. There, the basal erosion surface in both Fosses A and B describes a concave-up cross-sectional profile in a NE–SW orientation. Fosse A is an elongate, WNW–ESE-oriented depression that has a length of ~4 km and width of 0.9 km. The depression has a depth of ~80 m with flank slopes of up to 15°. The depression is eroded into Cretaceous strata (Figure 5.3) that form the hanging wall of a reverse fault, and progressively cuts through the Lower Chalk and Gault Clay into the Lower Greensand strata (Figure 5.5a; see also Chapter 4). The Fosse sediment infill comprises two distinct seismic units. The lowermost unit is characterized by a transparent seismic facies with weak parallel internal reflectors that drape and onlap the basal erosion surface. This lower infill is truncated by a distinct erosional surface (labelled IES) that indicates a second episode of scouring, before continued infilling. Fosse B is a ~2 km radius depression with a sub-circular planform that is eroded ~80 m through Lower Greensand sandstones into strata of the Wealden Beds (Figure 5.5a). The stratigraphy of this Fosse also comprises two distinct seismic units separated by a prominent internal erosion surface indicating that Fosse infilling was not continuous.

The spatially localized occurrence of the Fosses depressions is clearly demonstrated in Figure 5.6, which shows the geometry of Fosses D and E in consecutive NE–SW-oriented seismic profiles. Fosse E shows marked variability in both width and depth of erosion when traced from the northwest to southeast. By contrast, Fosse D is a ~90 m deep depression with a concave-up geometry in three-dimensions (Figure 5.7). Sub-horizontal strata within Fosse D onlap the basal erosion surface on all its flanks indicating it forms a closed depression.

5.3.3 Spatial distribution of the Fosses Dangeard

A map of sediment thicknesses within the Fosses (Figure 5.5b) reveals that they form spatially isolated sub-circular to elliptical features clustered in a ~7-km-wide, WNW–ESE-oriented belt, perpendicular to the Strait between Dover and Calais (Figure 5.3). This belt is parallel to the strike of Cretaceous strata occurring immediately south-west of the boundary between Lower Cretaceous bedrock and the Lower Chalk. Similar deep, bedrock-eroded, sediment-infilled depressions have not been reported either to the northeast or southwest of this narrow belt in the Strait. A fluvial valley interpretation is not tenable as they form closed three-dimensional (basin-shaped) erosional features, rather than two-dimensional (valley-shaped) features. Sets of the Fosses not only occur along the width of much of the Strait, but also occur in a NE–SW orientation parallel to the Strait, for example, Fosses A, B and C.

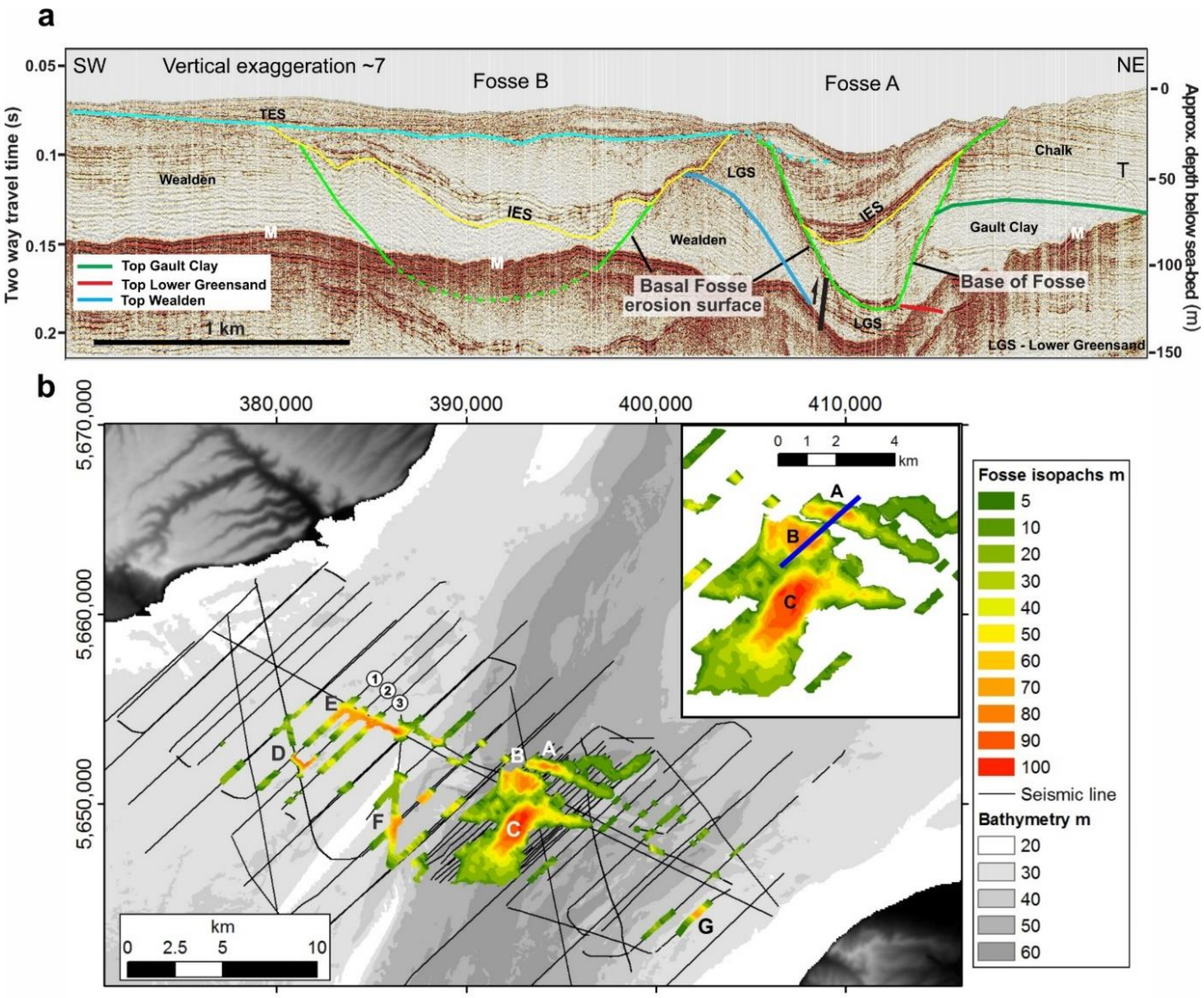


Figure 5.5. Morphology and distribution of the Fosses Dangeard. a) Interpreted seismic reflection profile across Fosses Dangeard depressions A and B showing cross-sectional geometry of Fosses and bedrock geology (location indicated in (b) by a blue line). IES, internal erosion surface within Fosse sedimentary infill; M: seabed multiple; T: seismic diffraction pattern caused by Channel Tunnel; TES: transgressive erosion surface. Location of seismic profile indicated in b inset. b) Map showing sediment thickness in Fosses Dangeard depressions. Inset shows detail of Fosses A, B and C. Note absence of color along seismic track-lines indicates bedrock exposed at seabed, indicating that the depressions are highly localized.

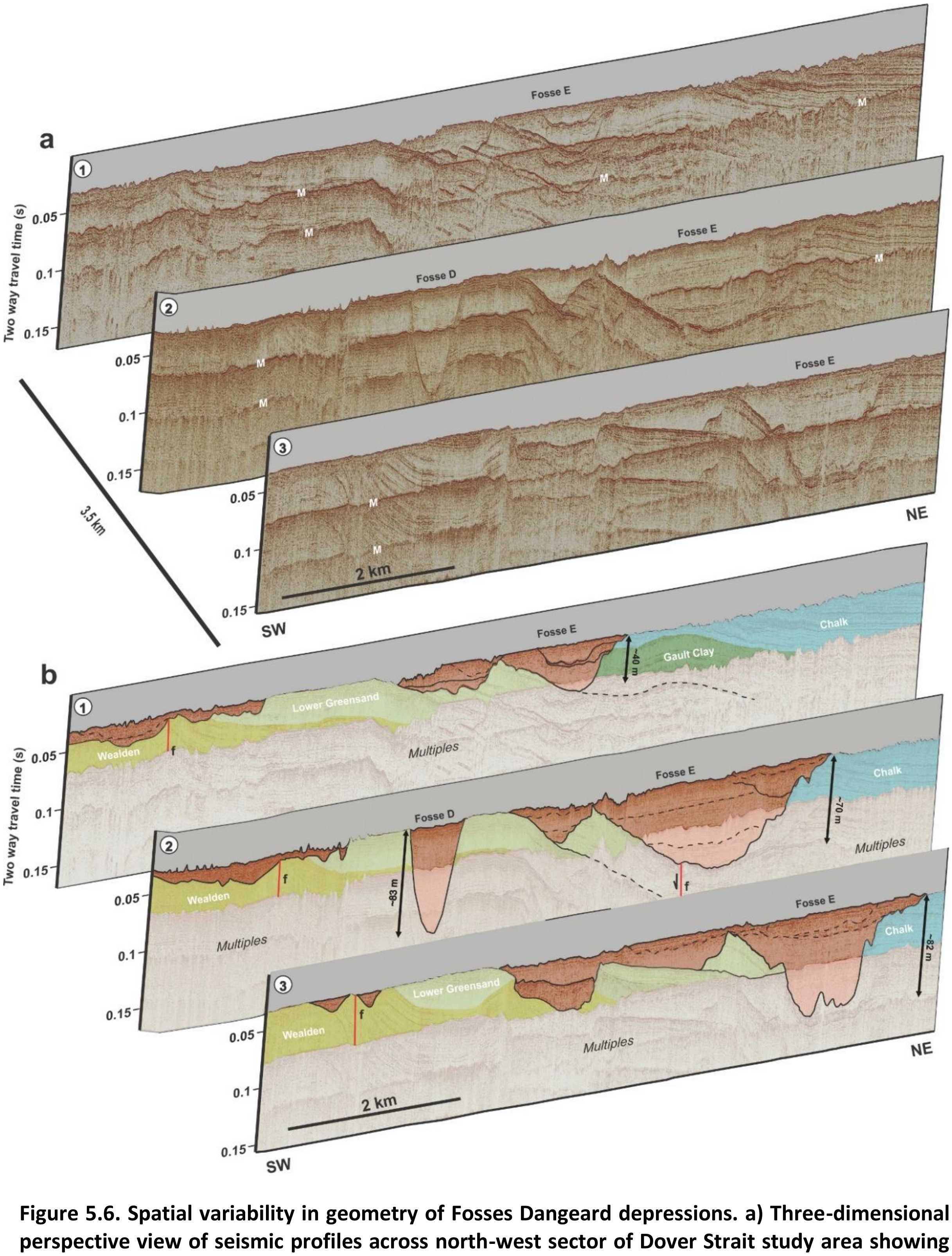


Figure 5.6. Spatial variability in geometry of Fosses Dangeard depressions. a) Three-dimensional perspective view of seismic profiles across north-west sector of Dover Strait study area showing geometry of depressions D and E. Depressions form localized features eroded in Cretaceous bedrock. Note presence of possible internal erosion surfaces within Fosses sediment infill. Vertical exaggeration ~13. M: seabed multiple. b) Geological interpretation of seismic profiles with details of Cretaceous bedrock into which Fosses are eroded. Locations of seismic profiles indicated in Figure 5.5b. f: fault.

5.3.4 Bedrock geology and structural influence on Fosses geometry.

The role of bedrock geology and pre-existing structure is an important consideration in characterizing the geometry of the Fosses Dangeard. The Fosses are eroded into a variety of lower Cretaceous bedrock lithologies with varying resistance to erosion. Fosses D and E are eroded up to 80 m into relatively more resistant Lower Greensand and Lower Chalk strata (Figure 5.6), whereas Fosses B, C and F are eroded into the less resistant Weald Clay (Figure 5.5). In the latter case, we find that the Fosses are relatively deeper and with a greater planform diameter than those eroded into more resistant lithologies. Thus, while the formation of the Fosses is not purely localized by differential erosion along weaker bedrock strata, bedrock lithology does influence their overall geometry. Similarly, pre-existing structures influence but do not govern Fosses formation. Fosse A is eroded along a WNW–ESE-oriented fault zone (see Chapter 4; also Van Vliet-Lanoë et al., 2010), which it cross-cuts and which constrains its geometry as indicated by its WNW–ESE-oriented elongate geometry. By contrast, Fosse D is eroded into horizontally stratified bedrock and is not associated with any structures (Figure 5.6 and Figure 5.7). Similarly, a seismic profile across Fosse B shows that while its north-eastern margin is eroded into the flank of a fault-controlled monocline, the overall geometry of Fosse B is not structurally controlled (Figure 5.5; see Chapter 4). We conclude that pre-existing structures, although influencing Fosses geometry, are not a primary control on their formation.

5.3.5 Geomorphic interpretation of the Fosses Dangeard

The plan-view and cross-sectional geometry of the Fosse depressions, together with their great depth of erosion and spatially localized occurrence leads us to interpret them as giant plunge pools, as originally speculated by Smith (1985). We propose that they formed by vertical drilling into bedrock by water jets derived from large waterfalls (Howard et al., 1994; Scheingross and Lamb, 2016; Lamb et al., 2007). Similar features are described albeit on a much smaller scale in natural examples (Howard et al., 1994; Lamb et al., 2007; Anton et al., 2015) and in laboratory experiments (Pagliara et al., 2008). The remarkable depth of erosion of these features into bedrock suggests that there must have been a substantial elevation drop adjacent to the scours to generate water jets capable of such erosion. It is also plausible that the plunge pools were formed as bedrock step-pool features by high-magnitude water flow down the steep SW-facing slope of the escarpment (Yokokwa et al., 2013; Scheingross, 2015). It is difficult to explain these giant depressions by either normal fluvial, glacial (Destombes et al., 1975; Kellaway et al., 1975) or tidal Hamblin et al., 1992) erosion processes.

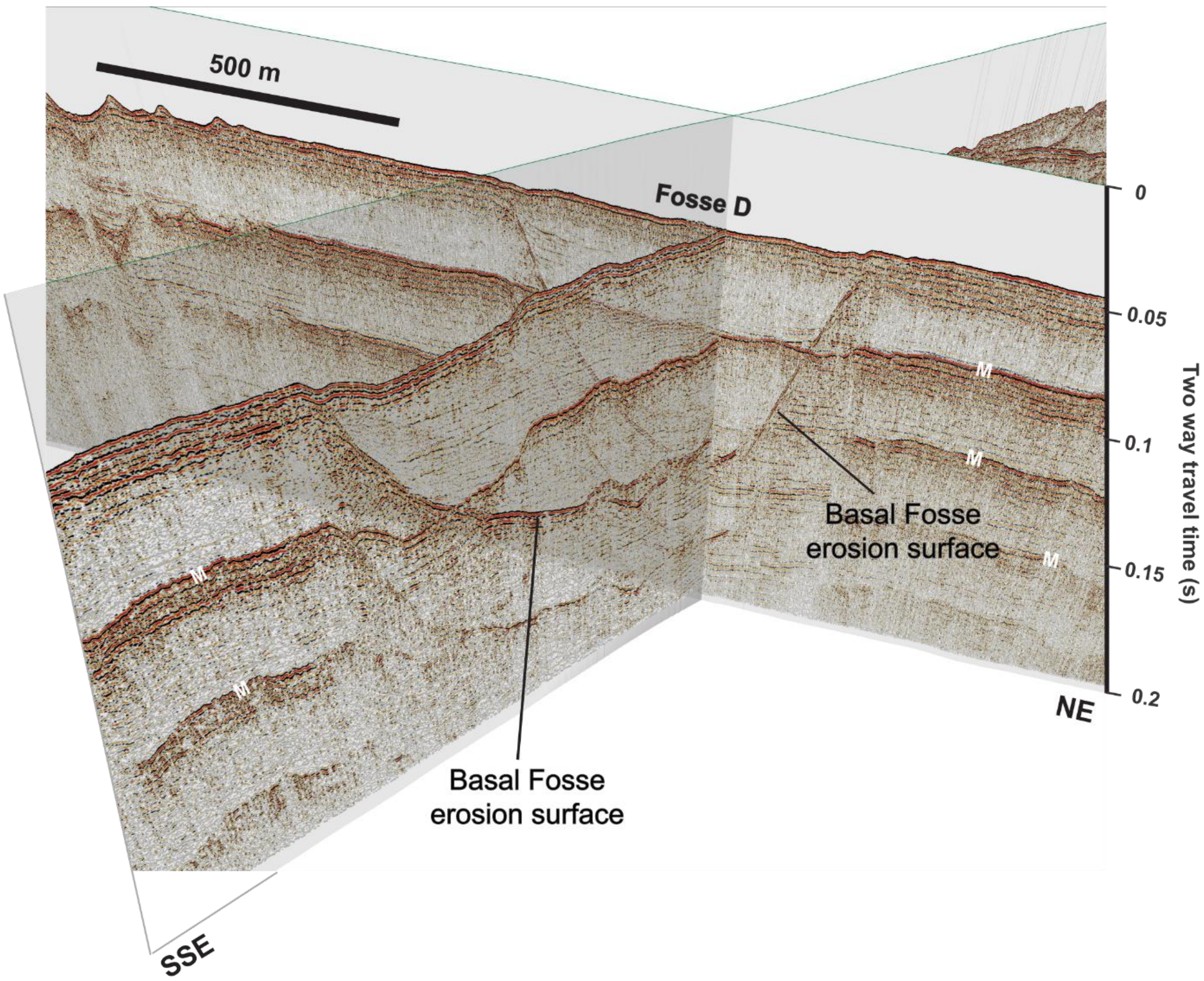


Figure 5.7. Three-dimensional perspective view of seismic reflection profiles across Fosse D (see location in Figure 5.5b). The NNW–SSE-oriented seismic line has been made partially transparent to view the geometry in NE–SW-oriented line. Note how the basal erosion surface of the Fosse, which is cut into Lower Cretaceous bedrock, shows a bowl-shaped geometry in crosscutting seismic profiles. Vertical exaggeration ~4. M: seabed multiple.

The plunge pool depressions are aligned with the offshore extension of a southwest-facing Chalk escarpment formed by the northeastern limb of the eroded Weald–Artois anticline (Figure 5.3 and Figure 5.4). The topographic profile of this escarpment onshore is markedly asymmetric with a shallow NE-facing dipslope and a steep, SW-facing escarpment (Figure 5.4b). Since this topographic feature is inferred to have formed the structural dam that impounded the southern North Sea proglacial lake at its southern limit (Gibbard, 1995; Smith, 1985; Gupta et al., 2007; Cohen et al., 2014), the presence of deeply eroded depressions at the foot of the escarpment strongly supports a model in which the plunge pools were formed by water cascading over this structural dam as large waterfalls. Lake water trapped to the northeast of the escarpment likely overspilled it to form large waterfalls that excavated the plunge pools. The escarpment retreated such that in the final stage the plunge pools eroded chalk bedrock. In this model, escarpment retreat by overspill would have produced sequential plunge pool erosion from Fosses C to A (Figure 5.5b). Similarly, in the northwestern part of the Strait, Fosse D may have been eroded when the escarpment was located at its northeastern margin. Subsequent, retreat of the escarpment ∼4 km to the northeast led to erosion of Fosse E. This relative timing of Fosses erosion cannot be constrained by our data. The infilling history of the Fosses remains poorly understood without lithological calibration. However, the seismic data reveal that after their excavation the plunge pools experienced multiple episodes of infilling and scouring. This likely relates to the complex multi-stage history of the area as it was repeatedly exposed and transgressed during subsequent glacial and interglacial cycles.

5.3.6 Seabed geomorphology of the Dover Strait

Our new compilation of marine geophysical data shows that the erosion and infilling of plunge pools at the base of the rock dam was not the only event that shaped the Dover Strait. A second event is interpreted on the basis of erosional landforms identified from analysis of high-resolution multibeam bathymetric data (∼2-m-grid-spacing) coupled with regional single-beam bathymetric data (see Chapter 3).

The data allow a new interpretation of the Lobourg Channel, which extends through the center of the Strait from the southern North Sea basin into the English Channel (Figure 5.4). The compilation shows the Lobourg Channel to be a ∼80-km-long and ∼10-km-wide NE–SW-oriented valley. It comprises a ∼25-m-deep valley eroded into Cretaceous bedrock, with a box-shaped cross-sectional profile (Figure 5.8a). Locally, lozenge-shaped bedrock remnants with streamlined margins are observed that we interpret as streamlined islands (Figure 5.8a), as identified in the downstream extension of the Lobourg Channel, the Northern palaeovalley in the central English Channel (Gupta et al., 2007; Collier et al., 2015).

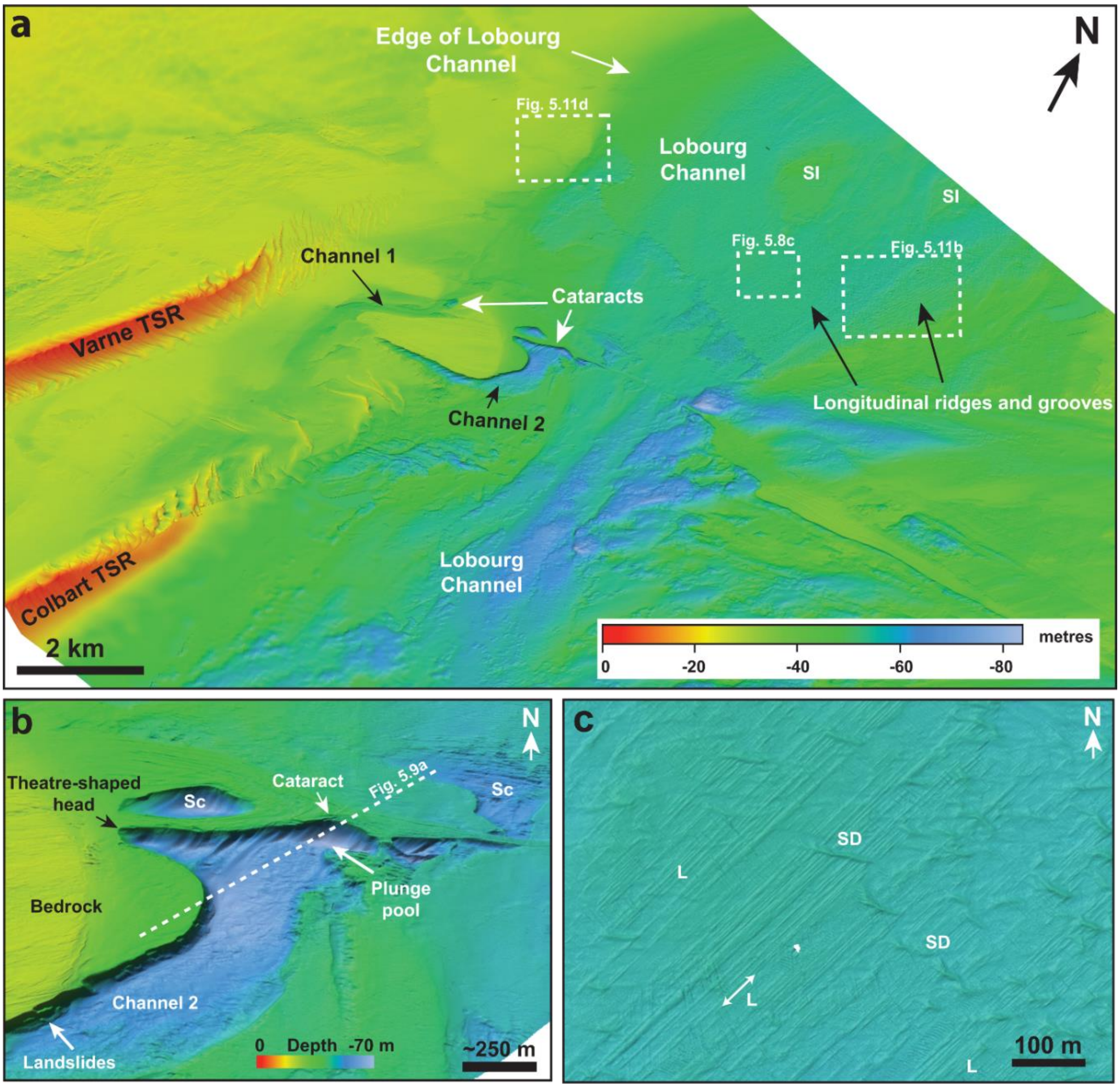


Figure 5.8. Sonar bathymetry of the central Dover Strait region. a) Colored and shaded relief multibeam bathymetry map of the Lobourg Channel. SI: streamlined island; TSR: tidal sand ridge. Water depth is indicated by color bar. Location of image is shown in Figure 5.4. b) Three-dimensional perspective view of cataract at head of Channel 2 looking N. Sc: prominent scours in bedrock. Vertical exaggeration is ~2. Water depth is indicated by color bar. Dashed line indicates line of seismic profile in Figure 5.9. (c) Map view of floor of Lobourg Channel showing prominent longitudinal lineations (ridges and grooves) (L). Orientation of lineations is indicated by double-headed arrow. SD: sand dunes. Location of image is indicated on (a).

The floor of the Lobourg Channel is itself locally incised by two narrow inner channels, ~500 m wide, that are linear to slightly sinuous in planform (Figure 5.8). Channel 1 is >3.5 km long and terminates to the northeast in an amphitheater-shaped head; it shows relief of ~15 m at the seafloor. It is eroded into relatively resistant sandstones of the Lower Greensand Formation. Channel 2 shows ~20 m of relief at the seabed, is ~200–500 m wide and branches to the northeast into two amphitheater-shaped heads. The inner channel headwall reaches a maximum of 24 m height, ~1 km wide and has a slope of ~10°–15° (Figure 5.8b). Seismic data reveal the channel and headwall are incised into bedrock, with the headwall formed of relatively resistant sandstones of the Lower Greensand Formation and the channel base in weaker claystones at the base of the Lower Greensand (Figure 5.9). A prominent erosional scour occurs in an alcove at the base of the headwall (Figure 5.9), which bathymetry data indicates is ~10 m deep. Immediately north of the headwall, a ~600-m-wide ovoid scour is eroded ~20 m into Lower Greensand bedrock (Figure 5.8b). The floor of the Lobourg Channel truncates sediment-infilled Fosses Dangeard depressions that underlie it, indicating that erosion of the Lobourg Channel post-dates Fosses incision and infilling (Figure 5.9).

We interpret the amphitheater-shaped channel heads as abandoned cataracts formed by focused knickpoint propagation similar to those formed by high-magnitude floods in the Channeled Scabland (Bretz et al., 1956; Baker and Nummedal, 1978), Snake River Plains (Lamb et al., 2014) and Iceland (Baynes et al., 2015a; Baynes et al., 2015b). Upstream migration of cataracts during flood flow carved the narrow, bedrock inner channels downstream of the amphitheater heads. The 10-m-deep erosional scour at the base of the headwall of Channel 2 is suggestive of plunge pool erosion by water plunging over the cataract lip (Figure 5.8b and Figure 5.9). Similar plunge pools occur in alcoves at the base of cataracts in the Channeled Scabland (Baker and Nummedal, 1978) and flood-eroded terrains in the Snake River Plains of Idaho (Lamb et al., 2014). Upstream retreat of the cataract to form Channel 2 may have been facilitated by the strong-over-weak bedrock stratigraphy and toppling failure in the Lower Greensand sandstone lithology (Lamb and Dietrich, 2009). The depth of erosion into bedrock at the seafloor associated with the cataracts is evidence of powerful erosional processes, although it should be noted that the magnitude of these is much smaller than the Fosses Dangeard depressions that we interpret as plunge pools.

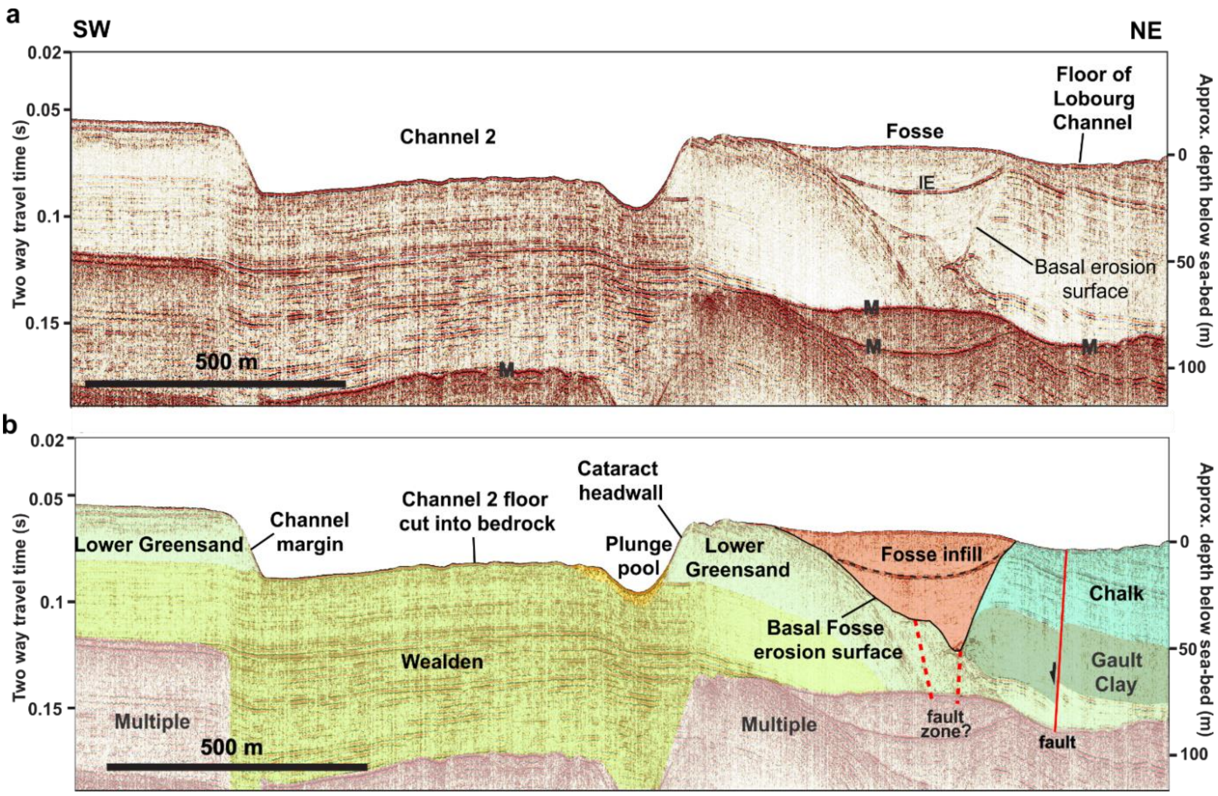


Figure 5.9. Seismic profile across inner Channel 2 and cataract headwall. a) Seismic reflection section and (b) geological interpretation. A prominent scour is present at the base of the headwall that we interpret as a plunge pool. Note that the floor of the Lobourg Channel upstream of cataract crosscuts a sediment-infilled Fosses Dangeard depression eroded into Cretaceous bedrock. Vertical exaggeration ~4. M: seabed multiple. Location of seismic profile is indicated in Figure 5.8b.

Upstream of the inner channels, we discovered several sets of parallel streamlined ridge-and-groove bedforms eroded into Chalk bedrock on the floor and immediate flank of the Lobourg Channel (Figure 5.8c, Figure 5.10 and Figure 5.11). These linear bedforms have a consistent ENE–WSW alignment (060°–240°), sub-parallel to the axis of the Lobourg Channel (Figure 5.10). Individual lineations are up to ~30 m wide, ~5 km in length, and have amplitudes of 0.5–1.2 m (Figure 5.11a). Locally sand dunes lie orthogonally across these features indicating that the lineations are not in equilibrium with the current flow field in the Strait. The origin of these highly parallel linear erosional bedforms remains enigmatic. Our preferred interpretation is that these represent bedrock-eroded longitudinal lineations developed by high-magnitude flood flows similar to erosional lineations observed in the Channeled Scabland of western Washington State, USA (Baker and Nummedal, 1978). Taken together, the presence of a linear bedrock-eroded channel with associated cataracts, deep scours and streamlined islands indicate that erosion of the Lobourg Channel was achieved by high-magnitude flood flows through the Strait.

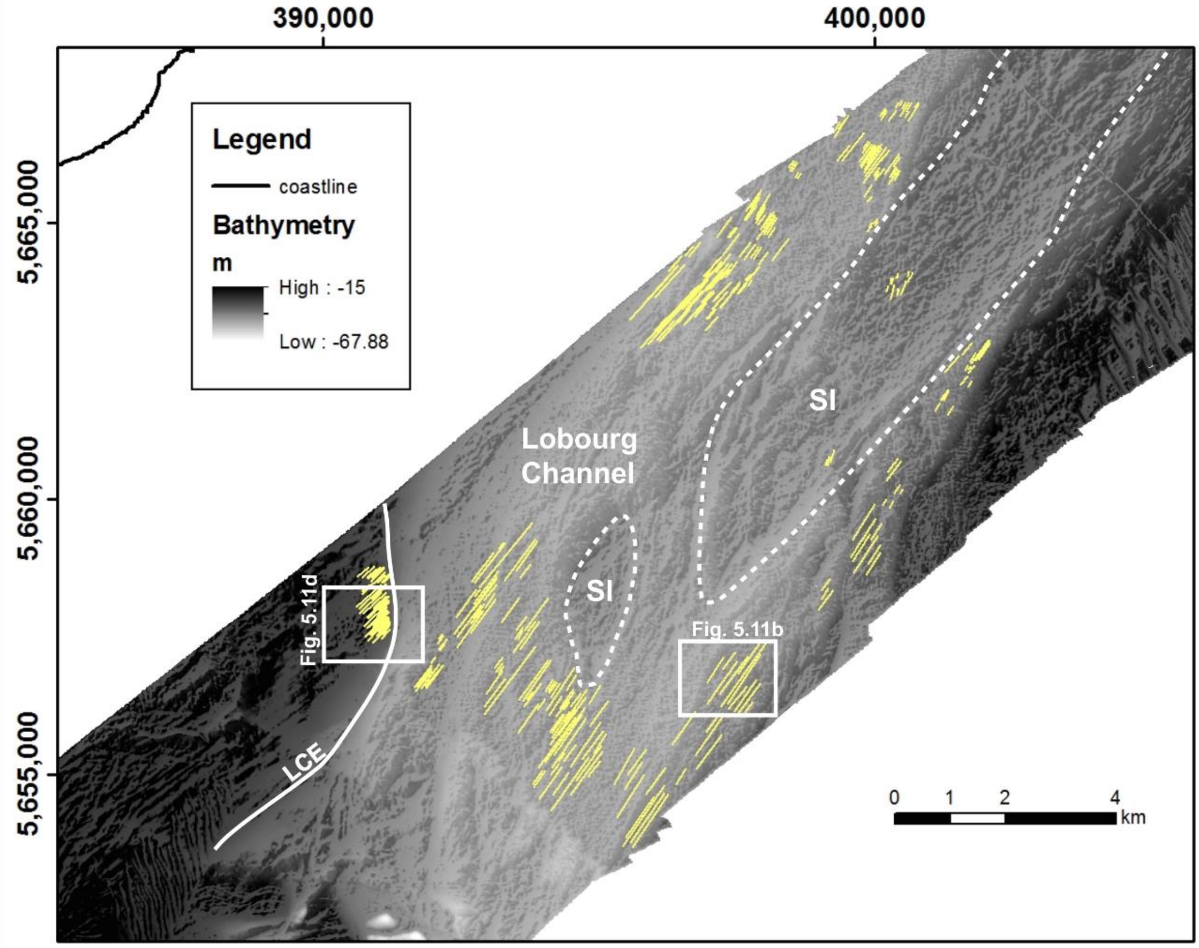
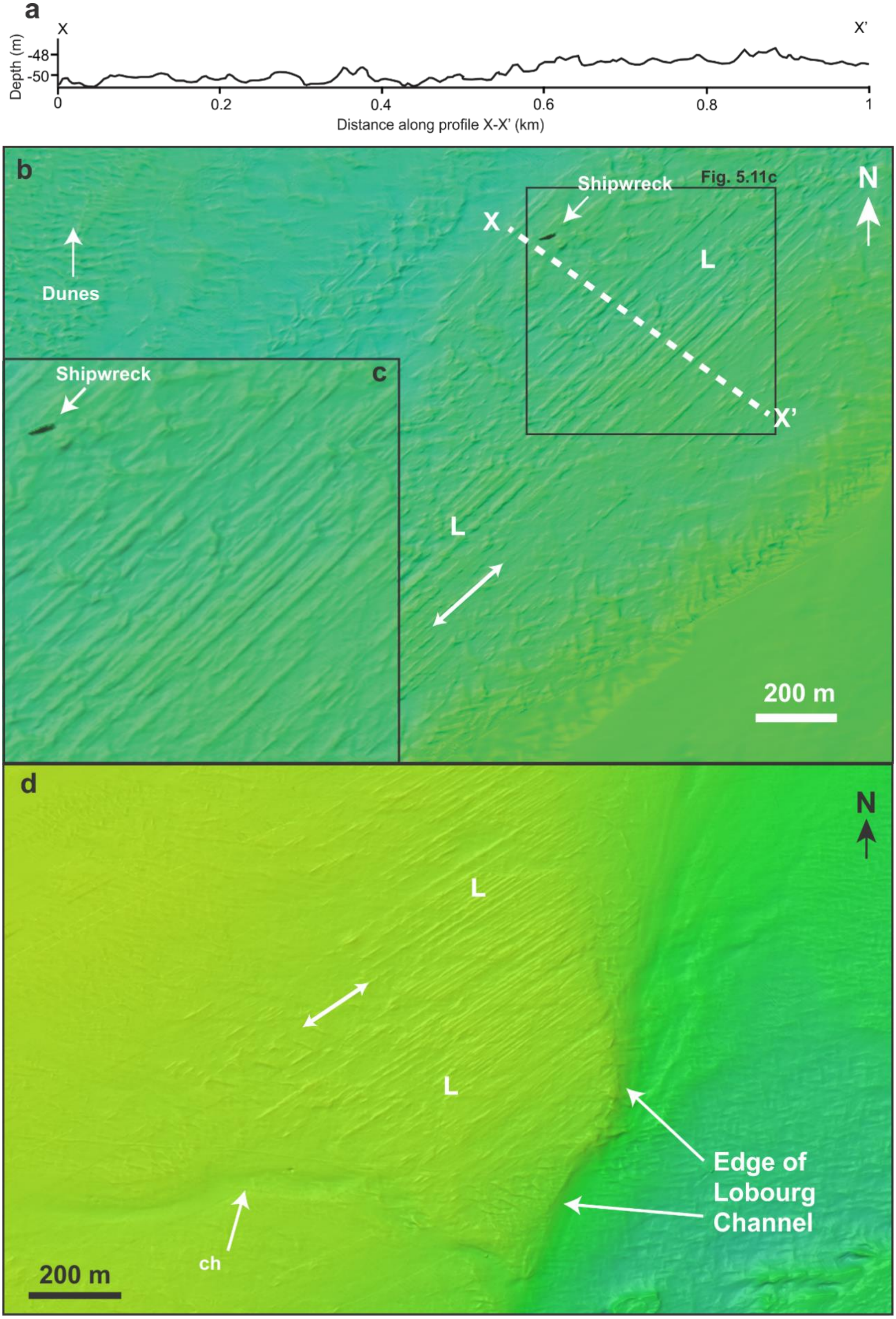


Figure 5.10. Map showing distribution and orientation of longitudinal lineations identified in multibeam bathymetric data from the Lobourg Channel. Longitudinal lineations are indicated in yellow. The background image is grey-scale multibeam bathymetry image. SI: streamlined island; LCE: Lobourg Channel edge. Locations of areas shown in Figure 5.11 are indicated. Note one set of lineations is eroded on north-west flank of Lobourg Channel.


a
X
X'
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-50
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0.2
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ch
200 m

Figure 5.11. Multibeam bathymetry images showing details of longitudinal lineations (L) eroded into Chalk bedrock. The orientation of lineations is shown by double-headed arrows. Location of images is indicated in Figure 5.8 and Figure 5.10. Water depth as in color bar in Figure 5.8. a) Topographic profile (X-X' in b) perpendicular to elongation direction of longitudinal lineations. b) Lineations at southeastern margin of Lobourg Channel. c) Enlarged area showing detail of bedforms. d) Lineations eroded into north-west flank of Lobourg Channel. Ch: small channel.

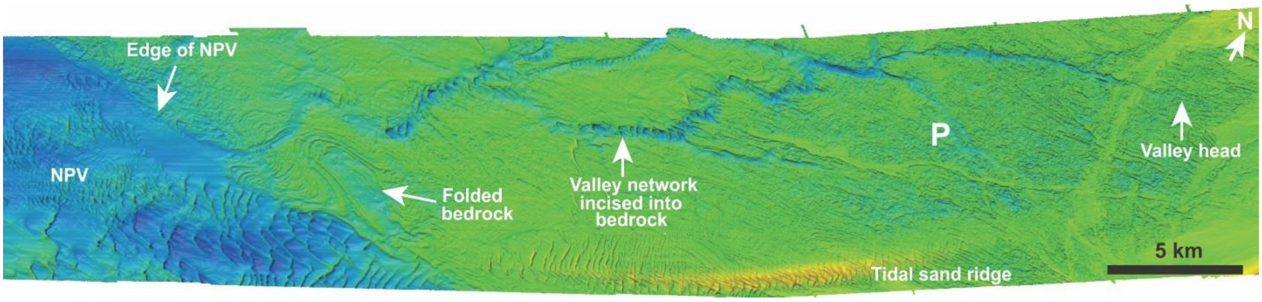

Figure 5.12. Multibeam bathymetry image showing detail of prominent valley eroded into bedrock of Platform P. Location indicated in Figure 5.4. Note how valley network heads on platform indicating that it is not connected to onshore drainage. NPV: Northern palaeovalley.

5.4 Discussion

Our marine geophysical data provide new insights into the late Quaternary landscape evolution of the Dover Strait with implications for understanding its history of breaching. In particular, the new data show that at least two significant erosional episodes shaped the opening of the Strait. While it is widely held that initial opening of the Strait was a consequence of spillover of a proglacial lake that existed in the southern North Sea basin during MIS 12 (Gibbard, 1988; Gibbard, 1995; Smith, 1985; Gupta et al., 2007; Cohen et al., 2014; Collier et al., 2015), direct evidence for this spillover process has up to now been lacking. The presence of large bedrock-eroded valleys with landforms associated with erosion by high-magnitude floods in the central and eastern English Channel has been proposed as evidence of catastrophic lake spillover (Gupta et al., 2007; Collier et al., 2015). Our detailed mapping of the geometry and distribution of the Fosses Dangeard depressions show that the Fosses represent plunge pool depressions formed at the base of a chalk escarpment. This provides compelling evidence of the presence of waterfalls from a proglacial lake dammed behind the Weald–Artois barrier. Spillover of lake water was clearly a key geomorphic process in the evolution of the Strait. The presence of multiple sets of depressions, some with an elongate geometry perpendicular to the strike of Cretaceous strata, suggests that erosion of the depressions may have been associated with retreat of the escarpment. A question that arises is the relationship of the Fosses Dangeard plunge pools to opening of the Strait. Did waterfall recession during lake overspill lead to eventual breaching of the escarpment and creation of the Strait? In this model, the plunge pools appear to be the only remnants of the breached dam. Given that the Fosses Dangeard depressions are distributed across much of the width of the Strait (Figure 5.3), together with their scale and depth of incision, it seems likely that there were several lake spillpoints across the rock ridge characterized by significant water overspill. The anabranching bedrock-incised valley network in the central and eastern English Channel may partly be a consequence of discharges resulting from breach of the rock ridge, but our data show that this interpretation is complicated by the effects of geomorphic overprinting by later events that eroded the Lobourg Channel. The presence of a prominent valley network some 20 km southwest of the Fosse, and eroded into bedrock platform P (Figure 5.4), may provide evidence of high-magnitude flows associated with breaching prior to Lobourg Channel erosion7. This ~40 km long network heads on the platform and shows no connection to onshore drainage indicating that it was initiated on the platform itself (Figure 5.12). In summary, we suggest that progressive recession of the chalk escarpment during lake overspill leading to eventual breaching of the dam provides a holistic model to explain initial opening of the Strait.

The combination of seismic reflection data and high-resolution bathymetry data reveals that a second event (or series of events) was needed to fully open the Dover Strait. The Lobourg Channel cross-cuts several of the Fosses Dangeard infills (for example, Figure 5.9), which clearly indicates that it formed as the result of younger erosional episodes. This distinction of two stages in the opening of the Strait is further substantiated by the fact that several sets of Fosse depressions lie outside the trace of the Lobourg Channel thus discounting any genetic link between the formations of these features (Figure 5.5b). The Lobourg Channel is characterized by geomorphic features indicative of erosion by catastrophic flood flows. The scale of the Lobourg Channel and presence of these characteristic landforms is suggestive that major episodes of flood flow were required to carve it.

Prior to formation of the Dover Strait, a prominent south-west-facing chalk escarpment extended across the Strait. The Channel region likely comprised a low-relief landscape during sea-level lowstands with a south-west flowing pre-breach Channel river fed primarily by river discharges from the Somme, Seine and palaeo-Solent rivers. Drainage from the escarpment to this river is likely to have been relatively small. We envisage the following sequence of events in the opening of the Strait itself: (1) spillover of pro-glacial lake water from multiple spillpoints along the chalk ridge led to excavation of several sets of plunge pool depressions, (2) retreat of the escarpment caused concomitant upstream propagation of Fosses erosion, (3) erosion of the ridge at multiple spillpoints causes eventual breaching of the Strait by rock dam failure with release of dammed lake water into the eastern English Channel, (4) multiple episodes of infill and scour of the Fosses during cycles of regression and transgression of the English Channel shelf, (5) incision of the Lobourg Channel and associated large valley network downstream by passage of a flood derived from the northeast in the North Sea basin. Erosion of the Lobourg Channel likely records the final opening of the Strait. The transition of the Lobourg Channel downstream into an anabranching network of bedrock valleys in the English Channel is suggestive of complex flood routing pathways southwest of the Dover Strait. We note that while events associated with lake overspill largely shaped the landscape of the Dover Strait, fluvial activity during lowstands with flow of a major river, the Channel River, through the Strait and marine erosion during high sea-level stands played a role in partially shaping the physiography.

While our results provide important new constraints on the relative timing of events that shaped the landscape evolution of the Dover Strait, our understanding is hampered by a lack of an accurate chronology for the sequence of events. Initial Pleistocene breaching of the Strait has been proposed to be an MIS 12 event (~450 ka and conventionally equated to the Elsterian–Anglian Stage glaciation). The extension of the Lobourg Channel tens of kilometers northeastwards from the Strait into the present-day southern North Sea basin (Figure 5.2) suggests that the flow(s) that carved this channel were likely derived from drainage of ice-marginal lakes in the central-southern North Sea basin (Cohen et al., 2014; Hijma et al., 2012; Busschers et al., 2008, Murton and Murton, 2012), or from flood flows derived from proglacial lakes further to the east (e.g. Meinsen et al., 2011). The timing of these events is not directly constrained, although a MIS 6 (~160 ka within the Saalian/Wollstonian Stage glaciation) is plausible. Marine molluscan fauna indicate that there was sporadic interconnection between the North Sea and English Channel during some of the high sea-level stands between MIS12 and MIS 6 (Meijer and Cleveringa, 2009; Hijma et al., 2012) suggesting partial breaching of the Strait had been accomplished during this time. By MIS 5e, marine mollusk assemblages from coastal deposits indicate full connection of the Channel with the North Sea during the interglacial highstand (Meijer and Preece, 1995). The two-stage evolution of the opening of the Dover Strait we propose is compatible with prior interpretation of at least two episodes of flooding from the central English Channel (Gupta et al., 2007; Collier et al., 2015), and sediment records from the Celtic Sea deep sea fans (Toucanne et al. 2009a, Toucanne et al. 2009b). Sediment delivery to these fans was pulsed with the greatest supply of sediment from the Channel routing system occurring during MIS stages 12 and 6 (Toucanne et al. 2009a, Toucanne et al. 2009b). Nevertheless, the lack of in situ derived chronologies from the Dover Strait (and from the English Channel more broadly) represents a hindrance to understanding its opening.

The Fosses Dangeard sediment infills are an outstanding target for future drilling in order to precisely constrain the chronology of events shaping the breaching history of the Strait, and its palaeogeographic consequences. Such a chronological framework is necessary to better understand the timing of when Britain first became isolated from mainland Europe during interglacial high sea-level phases. This has profound significance to understanding the ability and timing of biota, including humans, to colonize the British Isles (Busschers et al., 2008; Hijma et al., 2012). Furthermore, opening of the Strait caused large-scale re-routing of NW European drainage and meltwater to the North Atlantic via the English Channel (Toucanne et al., 2009a, Toucanne et al., 2009b). The re-routing of meltwater from the British-Scandinavian Ice Sheet and its injection into the North Atlantic has implications for inter-hemispheric climate variability (Toucanne et al., 2015, Plaza-Morlote et al., 2017).

References

Anton, L., Mather, A. E., Stokes, M., Munoz-Martin, A., and De Vicente, G., 2015, Exceptional river gorge formation from unexceptional floods. Nat. Communications, v. 6, 7963.

Arthur, J. C. R., Rankin, W. J. and Blundell, D. J., 1996, in Engineering Geology of the Channel Tunnel eds Harris C. S., Hart M. B., Varley P. M., Warren C. D. p. 129–143, Thomas Telford Publishing.

Ashton, N., and Lewis, S., 2002, Deserted Britain: declining populations in the British Late Middle Pleistocene, Antiquity, v. 76, p. 388–396.

Ashton, N., and Hosfield, R., 2010, Mapping the human record in the British early Palaeolithic: evidence from the Solent River system, Journal of Quaternary Sciences, v. 25, p. 737–753.

Ashton, N., Lewis, S. G., and Hosfield, R., 2011, Mapping the human record: population change in Britain during the Early Palaeolithic, p. 39-51. In: The ancient human occupation of Britain, (eds. Ashton, N., Lewis, J. E., and Stringer, C.).

Auffret, J. P., Alduc, D., Larsonneur, C. and Smith, A. J., 1980, Maps of the Paleovalleys and of the thickness of superficial sediments in the Eastern English-Channel. Annales De L'Institut Oceanographique, v. 56, p. 21–35.

Baker, V. R., and Nummedal, D., 1978, The Channeled Scabland; A Guide to the Geomorphology of the Columbia Basin, Washington, National Aeronautics and Space Administration.

Baynes, E. R. C., Attal, M., Dugmore, A. J., Kirstein, L. A., and Whaler, K. A., 2015, Catastrophic impact of extreme flood events on the morphology and evolution of the lower Jokulsa a Fjollum (northeast Iceland) during the Holocene, Geomorphology, v. 250, p. 422–436.

Baynes, E. R., Attal, M., Niedermann, S., Kirstein, L. A., Dugmore, A. J., and Naylor, M., 2015, Erosion during extreme flood events dominates Holocene canyon evolution in northeast Iceland, Proceedings of the National Academy of Sciences, v. 112(8), p. 2355–2360.

Bretz, J. H., Smith, H.T.U., and Neff, G. E., 1956, Channeled Scabland of Washington: New data and interpretations. Geological Society of America Bulletin, v. 67(8), p. 957–1049.

Busschers, F. S., Kasse, C., Van Balen, R. T., Vandenberghe, J., Cohen, K. M., Weerts, H. J. T., Wallinga, J., Johns, C., Cleveringa, P., and Bunnik, F. P. M., 2007, Late Pleistocene evolution of the Rhine-Meuse system in the southern North Sea basin: imprints of climate change, sea-level oscillation and glacio-isostacy, Quaternary Science Reviews, v. 26, p. 3216–3248.

Busschers, F. S., Van Balen, R. T., Cohen, K. M., Kasse, C., Weerts, H. J., Wallinga, J., and Bunnik, F. P., 2008, Response of the Rhine-Meuse fluvial system to Saalian ice-sheet dynamics. Boreas v. 37, p. 377–398.

Collier, J. S., Oggioni, F., Gupta, S., García-Moreno, D., Trentesaux, A., De Batist, M., 2015, Streamlined islands and the English Channel megaflood hypothesis, Global and Planetary Change, v. 135, p. 190–206.

Cohen, K.M., Gibbard, P.L., and Weerts, H.J.T., 2014, North Sea palaeogeographical reconstructions for the last 1 Ma, Netherlands Journal of Geosciences – Geologie en Mijnbouw, v. 93, p. 1–23.

Dingwall, R. G., 1975, Sub-bottom infilled channels in an area of Eastern English-Channel. Philosophical Transactions of the Royal Society of London A: Mathematical, Physical and Engineering Sciences, v. 279, p. 233–241.

Ehlers, J., Gibbard, P. L. and Hughes, P. D., 2011, Quaternary Glaciations-Extent and Chronology: a Closer Look, Elsevier.

Gibbard, P. L., 1988, The History of the Great Northwest European Rivers During the Past 3 Million Years, 1988, Philosophical Transactions of the Royal Society of London B: Biologic sciences., v. 318, p. 559–602.

Gibbard, P.L., 1995, The formation of the Strait of Dover, Geological Society, London, Special

Publications, v. 96, 15–26.

Gupta, S., Collier, J. S., Palmer-Felgate, A., and Potter, G., 2007, Catastrophic flooding origin of shelf valley systems in the English Channel. Nature, v.448, p. 342-345.

Hamblin, R.J.O., Crosby, A., Balson, P.S., Jones, S.M., Chadwick, R.A., Penn, I.E., and Arthur, M.J., 1992, United Kingdom offshore regional report: the geology of the English Channel. London: HMSO for the British Geological Survey, 106 p.

Hijma, M. P., Cohen, K. M., Roebroeks, W., Westerhoff, W. E., and Busschers, F. S., 2012, Pleistocene Rhine–Thames landscapes: geological background for hominin occupation of the southern North Sea region. Journal of Quaternary Science, v. 27, p. 17–39.

Howard, A. D., Dietrich, W. E., and Seidl, M. A., 1994, Modeling fluvial erosion on regional to continental scales. Journal of Geophysical Research: Solid Earth, v. 99(B7), p. 13971–13986.

James, J.W.C., Guennoc, P., Harrison, M., Le Bot, S., Philpott, S., Vinchon, C., Bee, E., Simien, F., Janjou, D., Garlan, T., Trentesaux, A., Mahieux, G., Briet D. and, Augris C., 2002, GEOSYNTH: A synthesis of the geology and sediments of the Dover Strait and its hinterland. CR/02/078, CDROM, British Geological Survey.

Kellaway, G. A., Redding, J. H., Shephard-Thorn, E. R. & Destombes, J. P., 1975, Quaternary history of English-Channel, Philosophical Transactions of the Royal Society of London A: Mathematical, Physical and Engineering Sciences, v. 279, p. 189–218.

Lamb, M. P., Howard, A. D., Dietrich, W. E., and Perron, J. T., 2007, Formation of amphitheater-headed valleys by waterfall erosion after large-scale slumping on Hawai ‘i. Geological Society of America Bulletin, v. 119(7-8), p. 805-822.

Lamb, M. P., and Dietrich, W. E., 2009, The persistence of waterfalls in fractured rock, Geological Society of America Bulletin, v. 121(7-8), p. 1123–1134.

Lamb, M. P., Mackey, B. H., and Farley, K. A., 2014, Amphitheater-headed canyons formed by megaflooding at Malad Gorge, Idaho. Proceedings of the National Academy of Sciences, v. 111, p. 57–62.

Lericolais, G., Auffret, J.-P., and Bourillet, J.-F., 2003, The Quaternary Channel River: seismic stratigraphy of its palaeovalleys and deeps, Journal of Quaternary Science, v. 18, p. 245–260.

Meijer, T. & Preece, R. C., 1995, Malacological evidence relating to the insularity of the British Isles during the Quaternary, Geological Society, London, Special Publications, v. 96, p. 89-110.

Meijer, T., and Cleveringa, P., 2009, Aminostratigraphy of Middle and Late Pleistocene deposits in The Netherlands and the southern part of the North Sea Basin, Global Planetary Change, v. 68, p. 326–345.

Meinsen, J., Winsemann, J., Weitkamp, A., Landmeyer, N., Lenz, A., and Dölling, M., 2011, Middle Pleistocene (Saalian) lake outburst floods in the Münsterland Embayment (NW Germany): impacts and magnitudes, Quaternary Science Reviews, v. 30, p. 2597–2625.

Mellett, C., Hodgson, M., Plater, A., Mauz, B., Selby, I. and Lang, A., 2013, Denudation of the continental shelf between Britain and France at the glacial–interglacial timescale, Geomorphology, v. 203, p. 79–96.

Murton, D. K., & Murton, J. B., 2012, Middle and Late Pleistocene glacial lakes of lowland Britain and the southern North Sea Basin, Quaternary International, v. 260, p. 115–142.

Pagliara, S., Hager, W. H., and Unger, J., 2008, Temporal evolution of plunge pool scour, Journal of Hydraulic Engineering, v. 134, p. 1630–1638.

Plaza-Morlote, M., Rey, D., Santos, J. F., Ribeiro, S., Heslop, D., Bernabeu, A., Mohamed, K.J., Rubio, V., and Martins, V., Southernmost evidence of large European Ice-Sheet derived freshwater discharges during the Heinrich Stadials of the last Glacial Period (Galician Interior Basin, Northwest Iberian Continental Margin), Earth and Planetary Science Letters, v. 457, p. 213–226.

Preece, R. C., 1995, Island Britain: a Quaternary Perspective, v. 96, Geological Society Special Publication.

Scheingross, J. S., 2015, Mechanics of Sediment Transport and Bedrock Erosion in Steep Landscapes, PhD thesis, California Institute of Technology.

Scheingross, J. S., and Lamb, M. P., 2016, Sediment transport through self-adjusting, bedrock-walled waterfall plunge pools. Journal of Geophysical Research: Earth Surface, v. 121, p. 939–963.

Smith, A.J., 1985, A catastrophic origin for the palaeovalley system of the eastern English Channel. Marine Geology, v. 64, p. 65–75.

Sutcliffe, A. J., 1995, Insularity of the British Isles 250 000–30 000 years ago: the mammalian, including human, evidence, Geological Society, London, Special Publications, v. 96, p. 127-140.

Stuart, A. J., 1995, Insularity and Quaternary vertebrate faunas in Britain and Ireland, Geological Society, London, Special Publications, v. 96, p. 111-125

Toucanne, S., Zaragosi, S., Bourillet, J. F., Gibbard, P. L., Eynaud, F., Giraudeau, J., Turon, J.L., Cremer, M., Cortijo, E., Martinez, P., and Rossignol, L., 2009a, A 1.2 Ma record of glaciation and fluvial discharge from the West European Atlantic margin. Quaternary Science Reviews, v. 28, p. 2974–2981.

Toucanne, S., Zaragosi, S., Bourillet, J.F., Cremer, M., Eynaud, F., Van Vliet-Lanoë, B., Penaud, A., Fontanier, C., Turon, J.L., Cortijo, E. and Gibbard, P.L., 2009b, Timing of massive 'Fleuve Manche' discharges over the last 350 kyr: insights into the European ice-sheet oscillations and the European drainage network from MIS 10 to 2, Quaternary Science Reviews, v. 28, p. 1238–1256.

Toucanne, S., Zaragosi, S., Bourillet, J. F., Marieu, V., Cremer, M., Kageyama, M., Van Vliet-Lanoë, B., Eynaud, F., Turon, J-L, and Gibbard, P. L., 2010, The first estimation of Fleuve Manche palaeoriver discharge during the last deglaciation: evidence for Fennoscandian ice sheet meltwater flow in the English Channel ca 20–18ka ago, Earth and Planetary Science Letters, v. 290, v. 459–473.

Toucanne, S., Soulet, G., Freslon, N., Jacinto, R. S., Dennielou, B., Zaragosi, S., Eynaud, F., Bourillet, J-F., and Bayon, G., 2015, Millennial-scale fluctuations of the European Ice Sheet at the end of the last glacial, and their potential impact on global climate, Quaternary Science Reviews, v. 123, p. 113-133.

Van Vliet-Lanoe, B., Gosselin, G., Mansy, J. L., Bourdillon, C., Meurisse-Fort, M., Henriet, J. P., Le Roy, P., and Trentesaux, A., 2010, A renewed Cenozoic story of the Strait of Dover. Annales de la Société Géologique du Nord, v. 17(II), p. 59–80.

Westaway, R. and Bridgland, D. R., 2010, Causes, consequences and chronology of large-magnitude palaeoflows in Middle and Late Pleistocene river systems of northwest Europe. Earth Surf. Processes Landforms 35, 1071–1094.

Yokokwa, M., Kotera, A., and Kyogoku, A., 2013, in Advances in River Sediment Research (eds. Fukuoka, S., Hajime Nakagawa, H., Sumi, T. & Zhang, H.), p. 629–633, CRC Press.

Chapter 6

Characterization of Middle–Late Pleistocene palaeo-landscapes in the Dover Strait and their implication for the opening of the Strait.

Chapter 6 – Characterization of Middle–Late Pleistocene palaeo-landscapes in the Dover Strait and their implication for the opening of the Strait.

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Abstract

One of the most significant palaeogeographic modification that occurred during the Pleistocene Epoch in northwestern Europe is the opening of the Dover Strait, which caused major rerouting of drainage systems at continental scale. This event has been associated with the formation of the tens-of-meter deep Fosses Dangeard, and with the hundreds-of-kilometer long Channel palaeovalley system. However, the processes that formed these features are still poorly constrained. In the previous Chapter of this dissertation, we presented geomorphological evidence pointing to plunge-pool erosion produced by waterfalls as the most likely cause for the incision of the Fosses Dangeard, and megaflood erosion as the possible origin of the Lobourg Channel of the Channel palaeovalleys. In the present chapter, we extend our previous investigations by describing and analyzing in detail the morphologies and infills of the Fosses Dangeard and Lobourg Channel in order to provide further evidence supporting the plunge-pool and megaflood hypotheses, as well as to better constrain how and when these features formed.

The detailed morphological analysis of the Fosses Dangeard indicates that these features were most likely carved by northward retreating waterfalls. Their morphologies are consistent with a sudden or rapid end of the waterfall phase. Our observations are therefore compatible with the hypothesis postulating that the Dover Strait opened due to a sudden breach of the ridge over which the waterfall spilled, generating a megaflood in the English Channel. The infills of the Fosses Dangeard indicate the occurrence of 5–6 significant scouring-and-infilling episodes in the center of the Strait in the span between the formation of the Fosses Dangeard and the last phase of valley incision along the Lobourg Channel. Two of the internal erosional surfaces identified in their infills exhibit morphologies similar to scours carved by high-energy river and/or by high-magnitude flood erosions. In the case of the Lobourg Channel, our geomorphological investigation has revealed three phases of valley incision also produced by intense fluvial erosion and/or high-magnitude flooding. The youngest valley incision truncates the uppermost infill of the Fosses Dangeard and it is associated with erosional features similar to those carved in megaflood-eroded terrains. Hence, we conclude that a megaflood event may have taken place once the Fosses Dangeard were fully infilled. Comparisons of the results of this investigation with palaeogeographic models suggest that that megaflood might have occurred during the Last Glacial Maximum.

Keywords: Geomorphology, Pleistocene, Paleogeography, Western Europe.

Author contributions: S.G., J.S.C., M.D.B. and T.C. conceived, designed and coordinated the study. D.G-M., J.S.C. and F.O. processed bathymetric data. D.G-M, K.VAN. (K. Vanneste) and W.V. designed, coordinated and conducted RV Belgica geophysical surveys with important contributions from K.VER. (K. Verbeeck), H.J. and A.T. K.VAN. processed RCMG–ROB and Dangeard I seismic-reflection data with contributions from D.G-M. A.T. provided Dangeard I seismic-reflection data. D.G-M. analyzed and interpreted the data with contributions from K.VAN., M.D.B., S.G. and J.S.C. All authors discussed the results. D.G-M wrote the paper. All authors reviewed the manuscript.

6.1 Introduction

The submarine landscapes of the English Channel and southern North Sea are characterized by hundreds-of-kilometers long interconnected palaeovalleys, tens-of-meter deep palaeo-depressions and kilometer-scale linear sand ridges (Figure 6.1; James et al., 2002; Antoine et al., 2003; Lericolais et al., 2013; Mellett et al., 2013; Collier et al., 2015). These erosional/depositional features are the result of several extreme climatic changes that occurred during the Quaternary Period. These involved a succession of glacial–interglacial cycles that caused drastic changes in sea level, episodes of isostatic depression and post-glacial isostatic rebound, and major rerouting of drainage systems at continental scale, which in itself may have impacted on ocean circulation and climate evolution (e.g. Zagwijn, 1989; Kjemperud and Fjeldskaar, 1992; Clark et al., 1999; Gibbard, 1995; Ehlers and Gibbard, 2004; Busschers et al., 2005; Busschers et al., 2007; Toucanne et al., 2009a; Toucanne et al., 2009b; Toucanne et al., 2010; Toucanne et al., 2015). Even though many studies have been undertaken in this area since the seventies, the formation and age of many of those geomorphological features are still under debate. Their characterization is indeed not straightforward, as the succession of changes of sea level, causing subaerial exposure and then submergence, may have erased and/or modified their original geometries and infills. In addition, little reliable dating is available, and high-resolution geophysical/geological data are still lacking from large parts of these areas.

Two of the most important and controversial erosional features located in the southern North Sea – Dover Strait area are the tens-of-meter deep, kilometer-scale Fosses Dangeard, and the palaeovalley known as the Lobourg Channel. Note that the latter forms part of the Channel palaeovalley system, also known as the Channel/Manche River (Figure 6.1; e.g. Mellett et al., 2013; Collier et al., 2015). The formation of those features is linked to one of the most significant geographical modification of northwestern Europe during the Pleistocene Epoch; i.e. the opening of the Dover Strait (Smith, 1985; Gibbard, 1995; Gupta et al., 2007; Gibbard and Cohen, 2015). The opening of the Dover Strait had indeed a significant impact on the organization of northwestern European hydrographic network and on biogeography, as well as on the pattern of early human colonization of Britain (e.g. Preece, 1995; Sutcliffe, 1995; Stuart, 1995; Meijer et al., 1995; Ashton and Lewis, 2002; Ashton and Hosfield, 2010; Ashton et al., 2011). Understanding the processes that formed the Fosses Dangeard and the Lobourg

Channel is therefore key to better reconstruct the Pleistocene history of northwestern Europe.

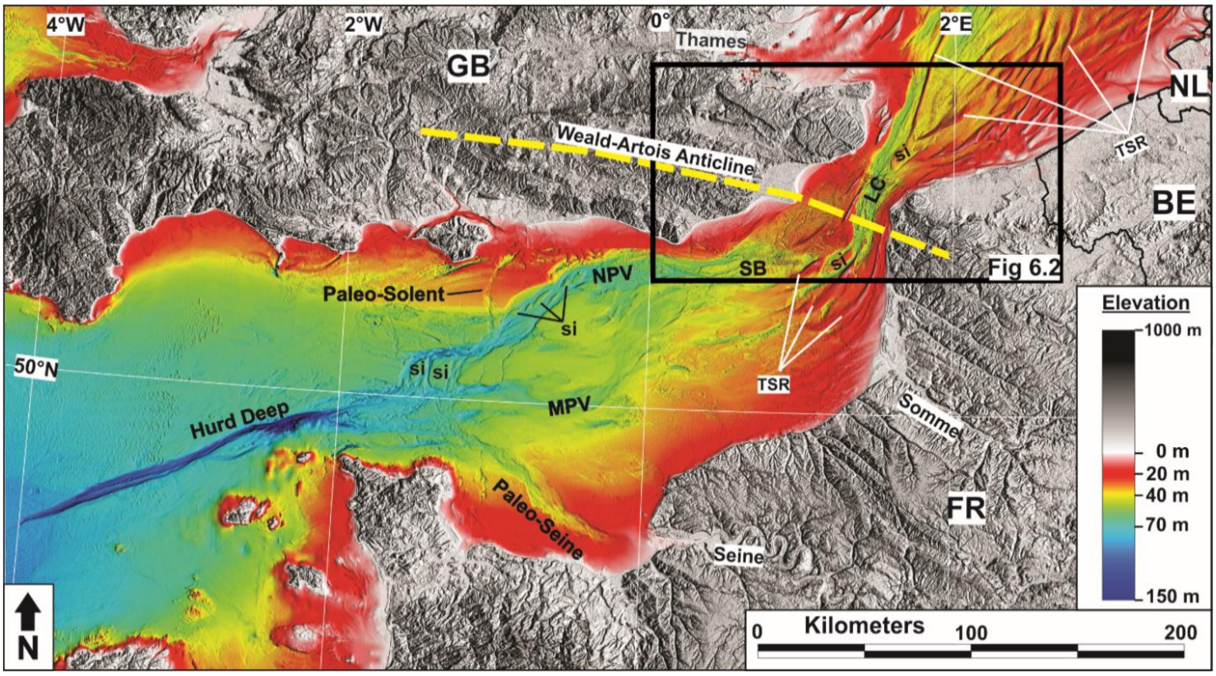

Figure 6.1. Bathymetric map of the English Channel. Bathymetric DTM produced by combining data from EMODnet (available at www.emodnet-bathymetry.eu) with those provided by UK and French hydrographic services (see Table 6.2). Topographic information: SRTM worldwide elevation data (3-arc-second resolution); available at www.cgiar-csi.org/data/srtm-90m-digital-elevation-database-v4-1. The various palaeovalleys composing the Channel palaeovalley system are indicated. These are: LC: Lobourg Channel; SB: South Basserelle palaeovalley; NPV: Northern palaeovalleys; MPV: Median palaeovalley. si: streamlined islands (see Collier et al., 2015); TSR: Tidal Sand Ridges. Black rectangle: area shown in Figure 6.2. For this and next figures, horizontal datum: WGS84; SRTM vertical datum: EGM96; Bathymetric vertical datum: Lowest Astronomic Tide (LAT).

According to palaeogeographic reconstructions (e.g. Gibbard, 1995; Gibbard and Cohen, 2015), northern France and southeastern Great Britain were connected through a land-bridge (typically referred to as "the Weald–Artois ridge"), which was formed by the northern limb of the Weald-Artois anticline. The Weald–Artois ridge appears to have separated the North Sea basin from the English Channel, at least, until ~450 ka BP (e.g. Gibbard, 1995). At that age, Earth was in the middle of the Elsterian glacial maximum, during which the Irish–British and Fennoscandian ice sheets merged for the first time across the central North Sea. The coalescence of the ice sheets caused the isolation of the unglaciated part of the southern North Sea from the North Atlantic Ocean, and blocked the northern drainage routes of palaeo-rivers (e.g. Gibbard, 1995; Murton and Murton, 2012; Gibbard and Cohen, 2015; Gibbard and Lewin 2016). That geographic setting, added to the isostatic depression possibly induced by the weight of the glacier, resulted in the formation of a proglacial lake in the unglaciated part of the southern North Sea. That lake was dammed at the Dover Strait by the Weald–Artois ridge and it may have extended tens-of-kilometer inland across Belgium and the Netherlands (e.g. Gibbard and Cohen, 2015; Gibbard and Lewin, 2016).

Several authors have theorized that, at some point during the Elsterian glacial maximum, the southern North Sea lake overtopped the Weald–Artois ridge, inducing the formation of waterfalls in the Dover Strait (e.g. Smith, 1985; Gibbard, 1995; Gibbard and Cohen, 2015). Even though evidence supporting this hypothesis is scarce, present consensus holds that these waterfalls led to the incision of the Fosses Dangeard due to plunge-pool erosion, and, eventually, to the opening of the Dover Strait.

Several studies have presented geomorphologic evidence suggesting that the formation of the Fosses Dangeard and Lobourg Channel may also be linked to the occurrence of megaflood events in the eastern English Channel (Smith, 1985; Gupta et al., 2007; Collier et al., 2015). This is consistent with the preliminary geomorphological investigation of the Dover Strait presented in Chapter 5 of this dissertation. The occurrence of high-magnitude flood flows in the English Channel is also supported by coarse gravels containing northern erratics and Fenno-Scandinavian heavy minerals found at the base of channels, as well as with massive fluvial discharges of the Channel palaeo-river that occurred during the Elsterian glacial maximum (Roep et al., 1975; Gibbard, 1995; Toucanne et al., 2009a). Several authors have therefore postulated that the opening of the Dover Strait was caused by a sudden breach of the Weald–Artois ridge, which generated a megaflood in the eastern English Channel that further incised the Fosses Dangeard and carved part of the Channel palaeovalleys (e.g. Smith, 1985; Gupta et al., 2007; Collier et al., 2015).

The sudden breach of the Weald–Artois ridge and the generation of a megaflood in the English Channel at Elsterian glacial maximum have not been demonstrated yet. Indeed, neither the erosional features nor the gravels suggesting catastrophic flooding identified in the Channel palaeovalleys have been linked to any specific glacial period owing to the lack of absolute age dating. In addition, the incision of the Lobourg Channel appears to postdate the formation of the Fosses Dangeard, suggesting the involvement of several major erosional episodes in the opening of the Dover Strait (see Chapter 5 of this dissertation). The erosional and sedimentary features indicating the occurrence of megafloods may hence have been formed during other marine lowstands. The occurrence of several intense fluvial and/or high-magnitude flood erosional episodes in the English Channel is also consistent with massive fluvial discharges along the Channel River that took place at Saalian and Weichselian glacial maxima (Toucanne et al., 2009a; Toucanne et al., 2010; Toucanne et al., 2015). In fact, palaeogeographic reconstructions propose that the ice sheets merged again across the southern North Sea during the Saalian Glacial maximum (MIS 6; 175–155 ka BP), inducing the formation of another proglacial lake in the southern North Sea (e.g. Gibbard, 1995). According to Gibbard and Cohen (2015), that lake was dammed in the north by the southern margin of the merged ice sheets and in the south by a land-bridge located a few kilometers to the north of the Dover Strait. The land-bridge damming the lake appears to have breached at some point before 150 ka BP, giving place to a major river system that comprised, among others, the Rhine, Meuse, Thames, Somme, Seine and Channel palaeo-rivers (Gibbard, 1988; Gibbard, 1995; Gibbard and Cohen, 2015). Hence, if that hypothesis is correct, the breach of the land-bridge and the emptying of the lake may also have induced episodes of high-magnitude flood flows and intense fluvial erosion in the Dover Strait during the Saalian glaciation. In addition, this area may have been reached by lake-outburst floods generated at other Saalian proglacial lakes located in the catchment of the palaeo-river system running through the Strait (e.g. Meinsen et al., 2011).

During the Last Glacial Period, the Fennoscandian and Irish–British ice sheets merged for the last time across the central North Sea between 30–19 ka BP (e.g. Sejrup et al., 2016; Carr et al., 2006). This caused the southward diversion of palaeo-rivers that ran into the unglaciated part of the southern North Sea, which converged again with the Channel palaeo-river in the Dover Strait (e.g. Gibbard, 1995). The resulted palaeo-river system had a mega-catchment area that exceeded 2.5 x 10^6 km^2, including the drainage of meltwater runoff from one quarter of the ice complex (Toucanne et al., 2015; Patton et al., 2017). The discharge of such a river system, sensitive to periodic meltwater injections triggered by ice sheet fluctuations, could indeed have induced intense fluvial erosion and episodes of flooding all along its extent.

The southern North Sea and English Channel were not emerged only during the glacial maxima of the various Middle–Late Pleistocene glaciations, but over almost the entire extent of these periods (i.e. between 478–424 ka; 374–130 ka; and 110–12 ka). Therefore, several other episodes of flooding and fluvial erosion may have contributed to shape the present-day Fosses Dangeard and the Channel palaeovalley system over the last 450 ka. Moreover, each glaciation was followed by an interglacial period. Three major marine highstands thus followed the Elsterian glaciation, reaching sea levels equal or slightly higher than the present-day (Meijer and Preece, 1995; Gibbard, 1995; Turner, 2000; Sturt et al., 2013). These are, the Holsteinian Interglacial (MIS 11; 400–350 ka BP), the Eemian interglacial (MIS 5e; 130–110 ka) and the Holocene Interglacial (MIS 1; 12,000 years ago – present). It is therefore quite likely that marine and coastal erosion significantly contributed to opened the Dover Strait and to shape/infill the Fosses Dangeard and Lobourg Channel.

In this chapter, we extend the investigation presented in Chapter 5 and in Collier et al. (2015) on the formation of the Lobourg Channel and Fosses Dangeard by describing and analyzing in detail the morphologies and infills of these erosional features. The main objective of the present study is to provide additional evidence demonstrating the plunge-pool erosion and possible subsequent megaflood(s) that incised the Fosses Dangeard, as well as to unravel the sequence of events that gave place to their present-day morphologies and infills. We also aim to better understand the erosional events that shaped the Lobourg Channel and their relationship with the formation and infilling of the Fosses Dangeard. In particular, we will firstly describe and interpret the 3-dimensional geometry and spatial distribution of the various palaeo-depressions composing the Fosses Dangeard. Based on that analysis, we will tentatively model the waterfall phase that incised those features. Second, we will analyze the infills of the Fosses Dangeard, including sedimentary facies and internal erosional surfaces, in order to establish the number and possible cause of the major erosional events that took place after the waterfall phase ended. Finally, we will perform detailed geomorphological analyses of the Lobourg Channel and its associated features to better characterize the erosional events that formed it. The combined interpretation of the infill and morphology of the Fosses Dangeard and Lobourg Channel will permit to assess the number and possible causes of the major erosional events (plunge-pool erosion, megafloods, river erosion, etc.) that shaped these features. Our findings will also allow the establishment of a more accurate relative chronology of the events that contributed to the opening of the Dover Strait. In addition, results from this study will shed light on the phases of evolution that led to the formation of the Fosses Dangeard.

6.2 Geological Setting

The seafloor of the Dover Strait exhibits a series of WNW-striking Cretaceous and Jurassic sedimentary formations that are locally covered by sand dunes and tens-of-meter to kilometer-scale sand ridges (Figure 6.1 and Figure 6.2; see also Hamblin et al., 1992; Dyer and Huntley, 1999; James et al., 2002; Reynaud et al., 2003). The Jurassic and Cretaceous geological formations are folded by the regional Weald-Artois anticline (Figure 6.2) and locally deformed/offset by the North Artois Shear Zone (see Chapter 4 of this dissertation). Quaternary deposits are rather localized in the Dover Strait, being mostly restricted to a few scattered Holocene sand ridges and dunes, and the infill the Fosses Dangeard (Figure 6.2 to Figure 6.4; see also James et al., 2002; Reynaud et al., 2003). The Lobourg Channel is mostly unfilled in the Dover Strait (see Chapter 4; James et al., 2002). This is tightly linked to the various submarine and subaerial erosional settings through which the Strait has passed during the Pleistocene Epoch, and the submarine erosional setting prevailing in this area during the Holocene (see Hamblin et al., 1992; Reynaud et al., 2003).

In this Chapter, we will refer to pre-Quaternary formations as bedrock. These units are extensively described in the literature from onshore exposures and offshore seismic data and boreholes (e.g. Crosby et al. 1988; Balson and D'Olier., 1989; Hamblin et al., 1992; James et al., 2002). Descriptions of the various geological formations outcropping in the Dover Strait are also provided in Chapter 4 of this dissertation. Anyhow, we list the main characteristics of the various sedimentary formations in Table 6.1.

The Chalk Group and the Lower Greensand sub-unit "Hythe Beds" represent the lithological units more resistant to erosion outcropping in the Dover Strait. They are well marked in the bathymetry by a series of ridges and scarps caused by differential erosion of hard and soft beds. In particular, the Hythe Beds stand out in the bathymetry as a 5–10 m high ridge that traverses the entire width of the Dover Strait (Figure 6.4).

Formations/ Groups	Subunits	Composition	Age	Reference
Chalk Group	Lower Chalk	Chalk (with flints nodules in the Upper Chalk Formation)	Late Cretaceous	Hamblin et al. (1992); Mortimore (2011)
	Middle Chalk			
	Upper Chalk			
Gault Clay/Upper Greensand		Clay and mudstones	Early Cretaceous	Hamblin et al. (1992); Hopson et al. (2008)
Lower Greensand	Atherfield Clay	Shales and mudstones	Early Cretaceous	Hamblin et al. (1992); Hopson et al. (2008)
	Hythe Beds	Alternating layers of "hard" sandy limestone and loosely cemented sandstones with glauconite.		
	Sandgate Beds	Argillaceous sands, sandstones and mudstones.		
	Folkestone Beds	Poorly consolidated sands with pebbles and clay.		
Wealden Beds	Weald Clay	Mudstones and shales	Early Cretaceous	Hamblin et al. (1992); Radley and Allen (2012)
	Hastings Beds	Interbedded sandstones, siltstones and shales		
Portland Beds and Purbeck Beds		Limestones, sandstones, mudstones and evaporites	Late Jurassic	Hamblin et al. (1992)
Kimmeridge Clay		Mudstones, sandstones and shales	Late Jurassic	Hamblin et al. (1992)

Table 6.1. Bedrock units outcropping in the Dover Strait.

The sediments infilling the Fosses Dangeard, their nature and absolute age, are poorly constrained. Only one sediment core exists from their infills (Destombes et al., 1975). This borehole sampled the first 50 m of the sediments infilling one of the palaeo-depressions composing the Fosses Dangeard (see Figure 6.4 for location). The sediments cored consisted of, from top to bottom, marine sands with pebbles, alternation of silty clay and fine–medium sand with traces of travertines, and a basal conglomerate (Destombes et al., 1975). Palynology suggests that those sediments were deposited during the Brørup interstadial (MIS 5c; ~96 ka) of the Weichselian glaciation. However, the lack of sediment cores from other locations within that or any other of the palaeo-depressions composing the Fosses Dangeard prevents the validation of that dating or the establishment of whether the sampled sediments represent local or widespread facies. Moreover, this dating may only represents the last phase of infilling of that palaeo-depression, since the core did not penetrate it entirely.

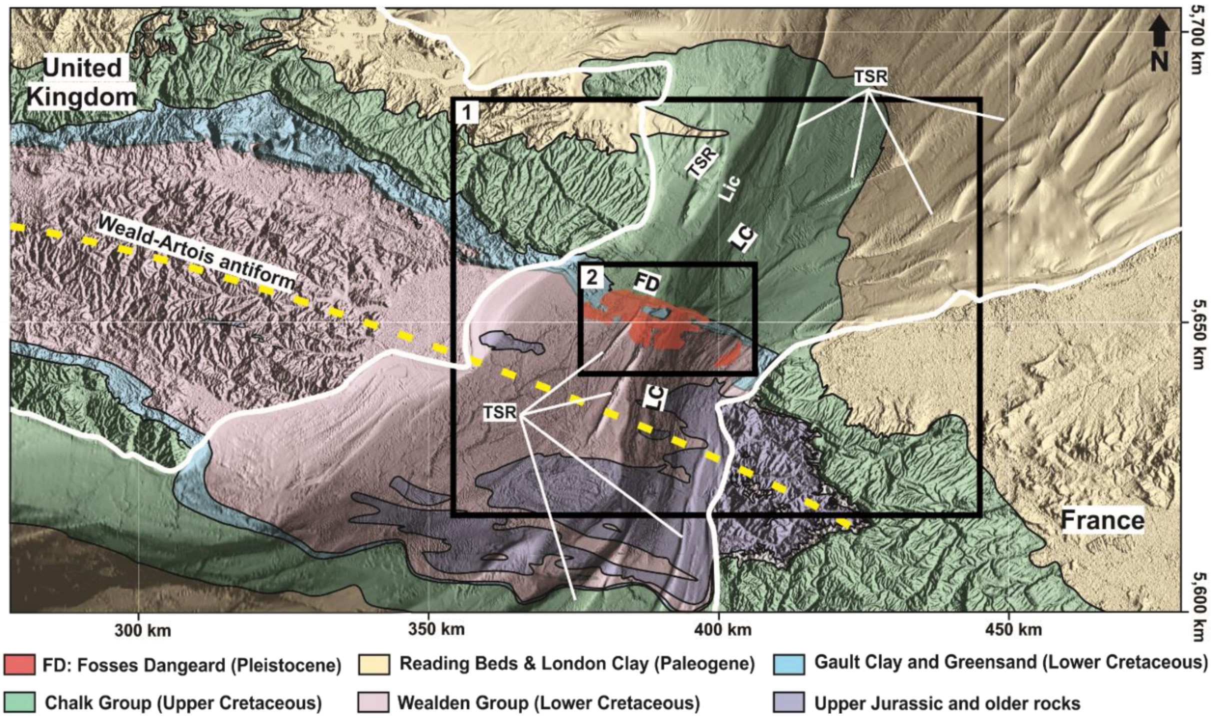

Figure 6.2. Simplified geological map superimposed to the bathymetric–topographic DEM of the Dover Strait region. After Balson et al. (1989) and Crosby et al. (1988). Location of Fosses Dangeard is indicated. Note that the Fosses Dangeard are carved along the northern limb of the Weald-Artois anticline. LC: Lobourg Channel; FD: Fosses Dangeard (red areas); Lic: Lobourg inner channel; TSR: Tidal Sand Ridges. Black square (1): areas shown in Figure 6.10; Black square (2): areas shown in Figure 6.4 and Figure 6.14. Projection: UTM, WGS 84, zone 31N (distances in kilometers).

The infills of some of the Fosses Dangeard are also interrupted by several internal erosional surfaces, attesting to the occurrence of various scouring-and-infilling episodes (See Chapters 4 and 5). The lowest sedimentary units infilling these palaeo-depressions may therefore be much older than the upper ones. Destombes et al. (1975) already suggested that the age yielded by the dating performed on the sampled sediments do not represent the time when the Fosses Dangeard formed. They proposed that these features were carved during the Saalian glaciation. More recent palaeogeographic reconstructions (see previous section of this chapter) have suggested though that the formation of these palaeo-depressions may be much older, associating it with erosional processes that occurred ~450 ka BP (e.g. Gibbard, 1995; Toucanne et al., 2009a; Gibbard and Cohen, 2015; Gupta et al., 2017). It is thus possible that the infill of these palaeo-depressions comprise sediments from the different marine highstands and lowstands that followed that age. However, reliable absolute dating is still lacking form these features. Hence, even though indirect sedimentary and geomorphologic evidence point to Elsterian glacial maximum, this age remains hypothetical.

6.3 Methodology

In the present study we combine 2D seismic-reflection data with single-beam and multibeam bathymetric data to interpret the 3D morphology, infill and interrelationship of the Fosses Dangeard and Lobourg Channel. The various datasets used in this study are a compilation of the datasets analyzed in Chapters 4 and 5 of this dissertation. Technical details on their acquisition and processing can be consulted in the Method section of this dissertation (see Chapter 3). In this chapter, we will therefore only summarize their main characteristics, which are listed in Table 6.2. The coverage of the various bathymetric datasets and seismic reflection profiles are shown in Figure 3.3 and Figure 4.4.

Data interpretation was carried out by combining 2D and 3D seismic interpretation software Opendtect and IHS Kingdom, with 3D topographic and geomorphologic analysis performed on GIS software Global Mapper and ArcMap. The combined interpretation of these datasets has resulted in detailed geomorphological maps of the Dover Strait's seafloor and isopach maps of the Fosses Dangeard more complete and accurate than those published in previous studies. Isopach maps were built by assuming mean seismic velocity through the palaeo-depressions' infills of 1,800 m s^{-1} (see Arthur et al., 1997). Only the infills of the Fosses Dangeard were considered to constrain the isopach maps. Major sandbanks and dunes were removed by subtracting the seafloor from the erosion surface at the base of these features.

Acquired/provided by	Year	Type of data	Processed at	Gridded at	V. resolution (m)	Max. Penetration (m)
Lille U. – RCMG	2002	SC	ROB	–	1–3	80–100
ROB–RCMG	2010–2012	SC & MC	ROB & RCMG	–	SC: 1–3; MC: 5–10	SC: 100–150 MC: ~250
ROB–RCMG	2010–2012	MBES	RCMG	5 m	–	–
MCA–UKHO	2006–2007	MBES	MCA – IC	1.5 m	–	–
UKHO	1988–2004	SBES	MCA – IC	30 m	–	–
SHOM	Since 1970's	SBES	Lille U.	40 m and 80 m	–	–
Emodnet	1946–2017	SBES & MBES	Multiple Institutions	230 m	–	–

Table 6.2. Geophysical datasets available for the present study. Lille U.: Lille University; RCMG: Renard Center of Marine Geology (Ghent University); ROB: Royal Observatory of Belgium; MCA: British Maritime & Coastguard Agency; UKHO: United Kingdom Hydrographic Office; SHOM: Service Hydrographique et Océanographique de la Marine; IC: Imperial College London; SC: Single-channel seismic-reflection data; MC: Multi-channel seismic-reflection data; SBES: single-beam bathymetric data; MBES: Multibeam bathymetric data.

6.4 The Fosses Dangeard

The basal erosional surface of the Fosses Dangeard is sharply mark in the seismic reflection data because of the angular unconformity that it forms and the strong differences between the bedrock seismic facies and that of the palaeo-depressions infill (Figure 6.3 to Figure 6.9). On some of the lines, the basal surface is below the first seabed multiple, yet it could still be mapped with confidence.

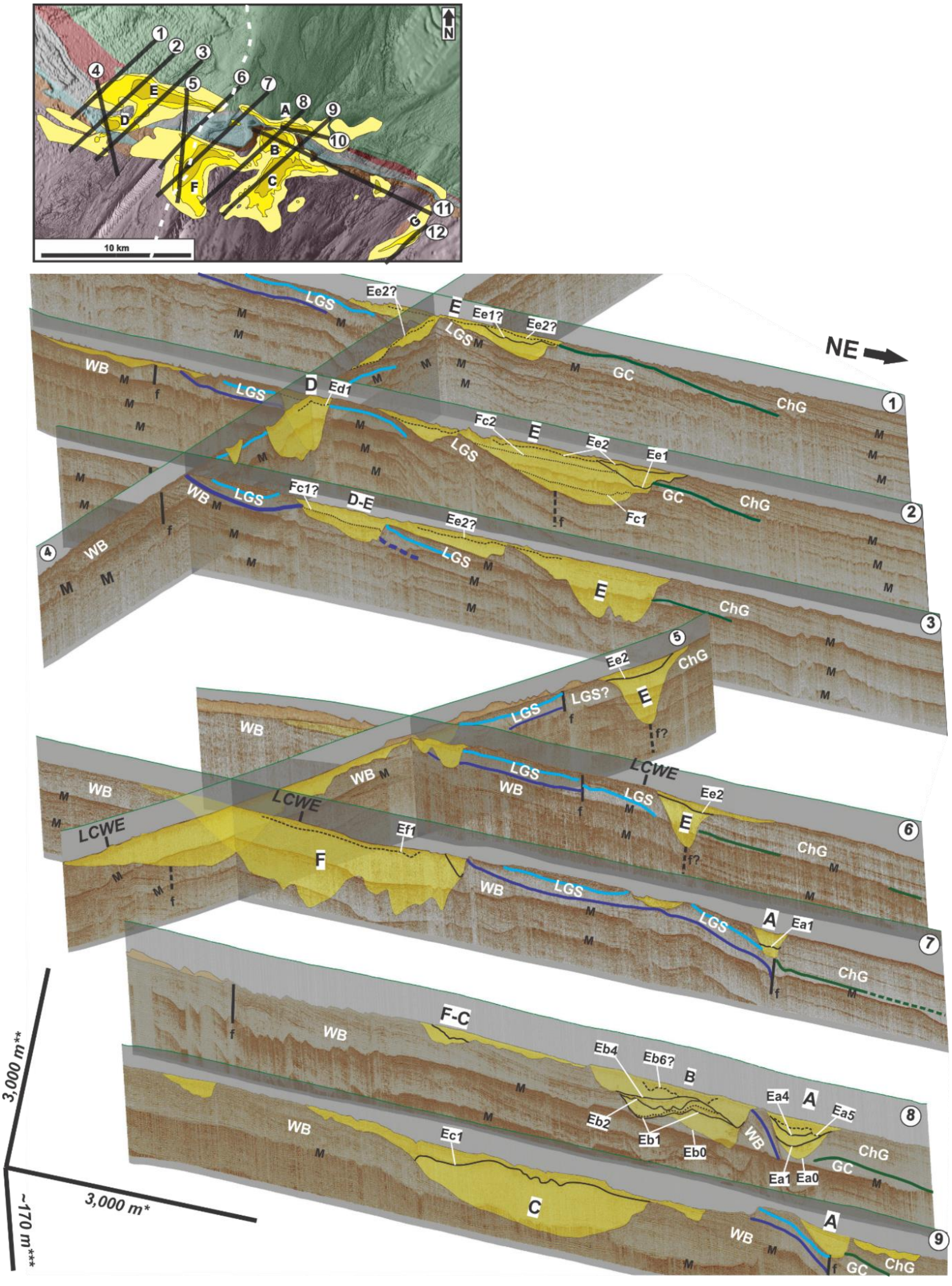

Figure 6.3. 3D plot of 9 interpreted seismic reflection profiles traversing the Fosses Dangeard (see location in Inset map). Yellow color: Fosses Dangeard's infill; black continuous and dashed lines: internal erosional surfaces (Eb1, etc.); black dotted lines: seismic horizons marking changes in seismic facies (e.g. Fc1 and Fc2); dark blue line: boundary Lower Greensand (LGS) – Wealden Beds

(WB) Formations; celeste blue line: Boundary Hythe Beds – Atherfield Clay; dark green line: boundary Gault Clay (GC) – Chalk Group (ChG) Formations; M: seabed repetition (i.e. seismic multiple); f: fault. LCWE marks the position of the western edge of the Lobourg Channel. *Distance measured along the orientation of the seismic profiles; ** distance measured across the orientation of the SW–NE seismic profiles; ***depth below the seafloor surface calculated assuming mean sound velocities of 2000 ms^{-1}.

Even though it has several blank areas due to local lack of data, the isopach map resulting from the mapping of the basal erosional surface of the Fosses Dangeard permits the interpretation of its morphology (Figure 6.4b). On the basis of the mapping, we have identified 7 discrete depressions that we named Fosse A–G. Their infills have internal erosional surfaces that we label Ea1, Ea2, Eb1, Eb2, etc., where "a" refers to internal erosions mapped in Fosse A, "b" to those mapped in Fosse B, etc., and the numbers are in order from lowest to highest. Prominent reflectors marking changes of seismic facies within the same seismic unit are also highlighted in the figures. We label these surfaces Fc1, Fc2, etc., where the numbers are in order from lowest to highest. Scours carved in the seafloor within the Lobourg Channel are labelled sc1, sc2 and sc3, where numbers refer to the erosional phase within which they were carved; e.g. sc1 indicates that the formation of that scour was likely linked to the valley incision represented by escarpment E1 (see section 6.5 of this manuscript). Labels d1 and d2 designate unfilled or partially infilled depressions carved in the seafloor, and Ch1 and Ch2 refers to prominent inner channels carved within the Lobourg Channel (see Chapter 5).

The Fosses Dangeard consist of several interconnected sediment-filled palaeo-depressions incised in the central part of the Dover Strait and distributed over a WNW–ESE elongated area of ~312 km² (Figure 6.4b). These features appear to be isolated (Figure 6.3), showing no connection with buried palaeovalleys located in the vicinity. The spatial distribution of the Fosses Dangeard appears to be independent from the Lobourg Channel. That is, three of the deepest Fosses are carved partially (Fosse F) or entirely (Fosses D and E) outside the Lobourg Channel (Figure 6.4). More importantly, the Lobourg Channel and its associated features (i.e. scours, etc.) truncate the uppermost infill of the Fosses located within it (e.g. Figure 6.5 and Figure 6.6).

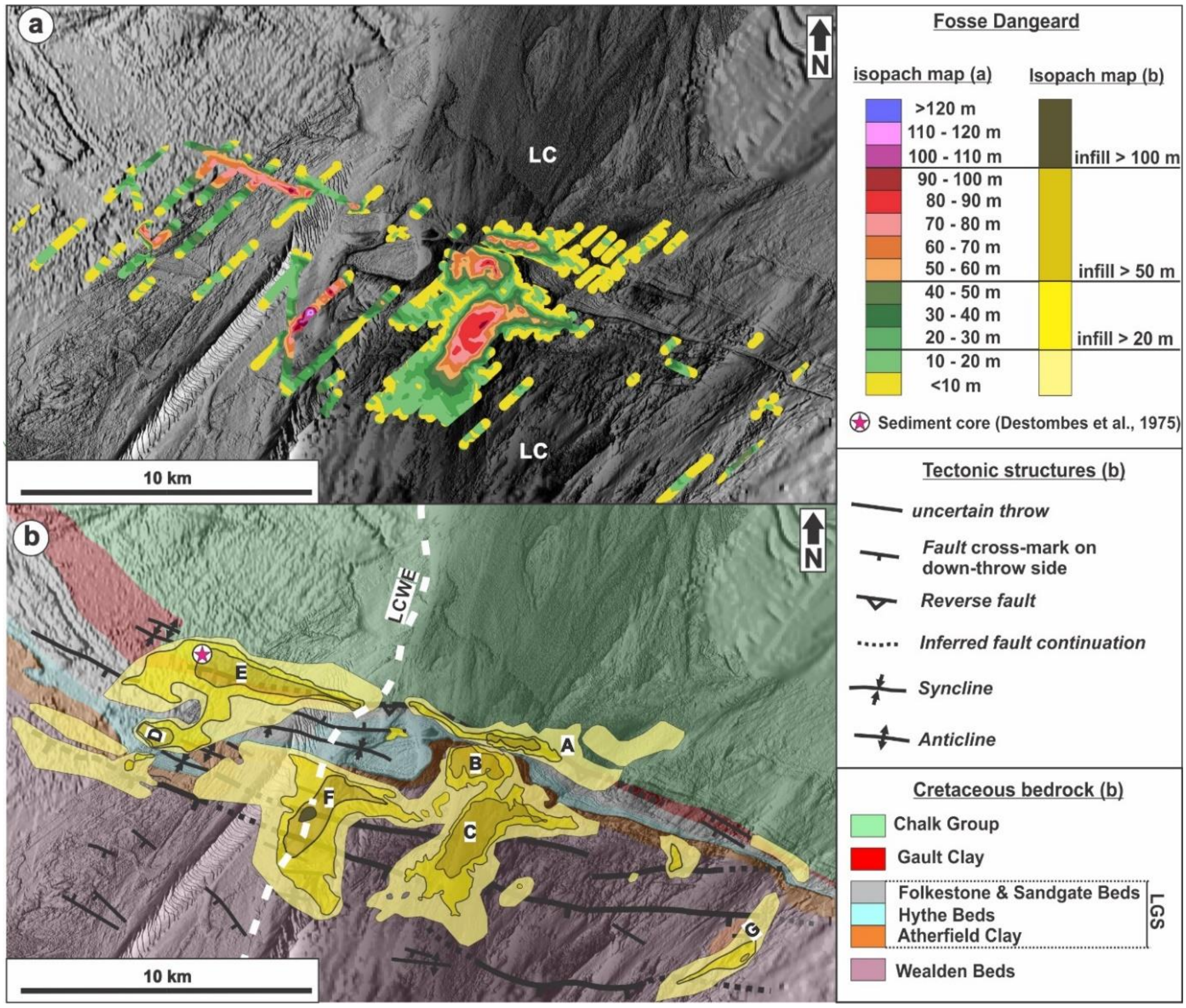

Figure 6.4. Geology of the central Dover Strait (see also Figure 6.14). a) Isopach map of the Fosses Dangeard superimposed to bathymetric merged, modified from Chapters 4 and 5. c) Geological/structural map of the central Dover Strait (see Chapter 4), including a morphological interpretation of the Fosses Dangeard based on the seismic reflection data (Figure 6.3 to Figure 6.9). Note that Fosse E, Fosse D, and part of Fosse F are carved outside the Lobourg Channel. LCWE: Lobourg Channel western edge; LC: Lobourg Channel.

The Fosses Dangeard comprise 7 major palaeo-depressions with maximum depths ranging between 50 m and 120–140 m, and several scattered minor sediment-filled incisions with maxima depths ≤ 20 m (Figure 6.4b). The various Fosses show a range of basal erosional geometries and sediment infill patterns (Figure 6.3). Their infills are interrupted by internal erosional surfaces, attesting to several scouring-and-infilling episodes. The number of internal erosions differs from one depression to another, even among interconnected ones, such as Fosses B, C and F (see Figure 6.6). Fosses A and B seems to have been subjected to many more scouring-and-infilling episodes than the rest (Figure 6.3). Here below, we describe the morphology and infill of each palaeo-depression in detail.

6.4.1 Fosse A

Fosse A is a 300–800 m wide, 70–80 m deep, E–W-elongated sediment-filled depression. This palaeo-depression is incised subparallel to the strike of the northern chalk limb of the Weald-Artois anticline, and it is entirely located within the Lobourg Channel (Figure 6.4). Longitudinal scours and an isolated depression (d1) possibly associated with the formation of the Lobourg Channel are clearly incised in its upper infill (Figure 6.5).

In cross section, the basal erosional surface of Fosse A displays a scoop-shape morphology perpendicularly to its major axis (Figure 6.3). This palaeo-depression cuts through the stratigraphic units Lower Chalk, Gault Clay, Folkestone Beds, Sandgate Beds and Hythe Beds. Fosse A is confined between the folded (due to the reverse faulting) Hythe Beds and the sub-horizontally stratified Chalk Group (Figure 6.3). In fact, it is mostly carved along the major reverse fault confronting these two stratigraphic units in the center of the Strait (Figure 6.4b). In addition, large parts of it are carved into the less resistant stratigraphic units Sandgate Beds, Folkestone Beds and Gault Clay. Hence, differential erosion due to the local geology and structure may have a role in determining Fosse A's morphology. However, the geological/structural control was probably only one of the factors, as the northern slope of Fosse A is incised into sub-horizontally stratified chalk units (Figure 6.3).

The sedimentary infill of Fosse A is not continuous. Seismic units within this palaeo-depression are interrupted by internal erosional surfaces (Figure 6.3 and 6). This attests to the occurrence of several scouring-and-infilling episodes following the initial bedrock incision. We have identified 5 major internal erosions, and we have labeled them, from the basal erosional surface (Ea0) upwards: Ea1–5. Erosion Ea1 extends over the whole of Fosse A, completely eroding locally the sedimentary package between Ea0 and Ea1 (Figure 6.5). The rest of the internal erosions are, on the other hand, more localized. In general, all internal erosions show cross-sectional scoop-shape morphologies across the width of Fosse A, and irregular and more random morphologies along its major axis (Figure 6.3 and Figure 6.5).

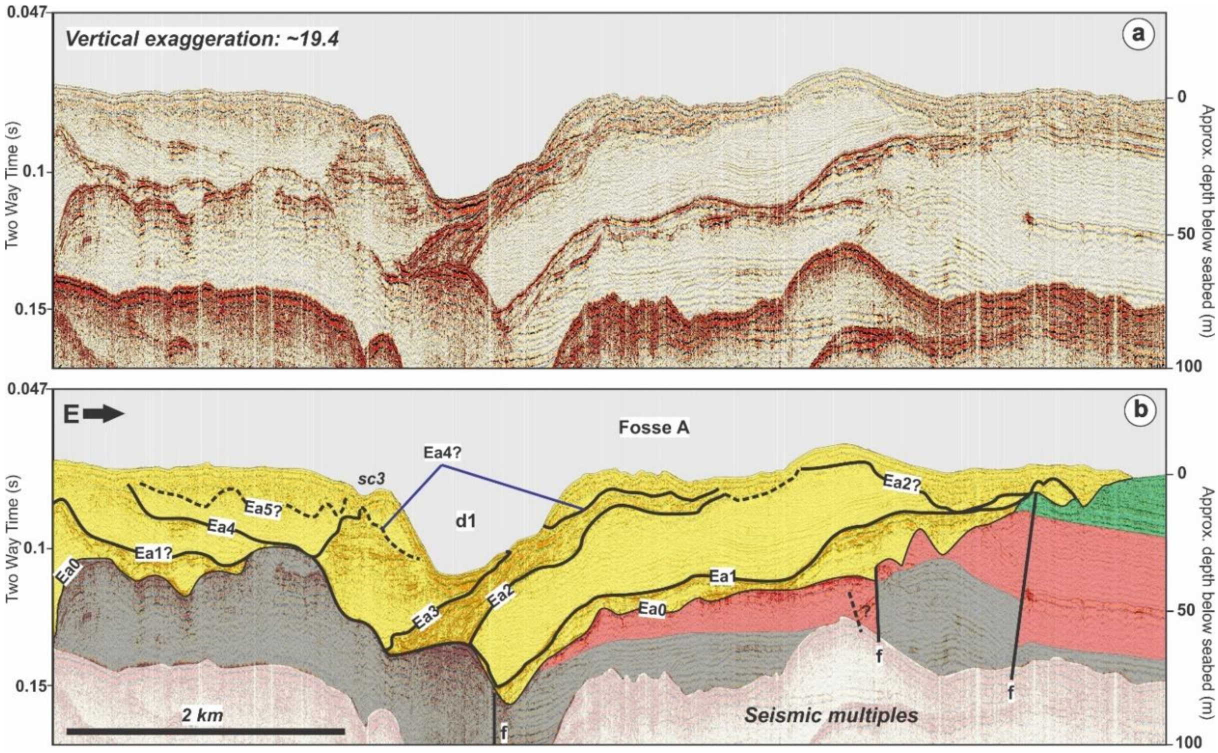

Figure 6.5. Non-interpreted (a) and interpreted (b) high-resolution single-channel seismic-reflection profile (profile number 10 in Figure 6.3 inset) acquired along Fosse A main axis. Yellow area: Fosse's infill; other colored areas: bedrock geology (see Figure 6.4); f: reverse fault crossing the seismic profile at 3 different locations; black lines within Fosse A: internal erosional surfaces; sc3: longitudinal scour (see Figure 6.14); d1: depression excavated into Fosse A's infill. For this and next interpreted seismic profiles, depth below seabed are calculated assuming a mean seismic velocity of 2,000 ms^{-1}.

The seismic facies among the various erosional surfaces can be classified in four groups:

- Ea0–Ea1 and Ea3–Ea4 infills: seismically transparent subunits that gradually evolve upwards to subunits with moderate-amplitude reflections;
- Ea1–Ea2 and Ea4–Ea5 infills: Low amplitude reflections to seismically transparent.
- Ea2–Ea3 infill: well-defined, subparallel high-amplitude reflections;
- Ea5–seafloor: diffused reflections.

6.4.2 Fosse B

Fosse B is the northernmost and shallowest (~80 m deep) sediment-infilled depression of a larger (~40 km²) incision that also comprises Fosses C and F (Figure 6.4b). Fosse B is mainly carved into the Weald Clay unit of the Wealden Beds. This depression is also located entirely within the Lobourg Channel (Figure 6.4). Longitudinal scours associated with the erosion of the Lobourg Channel (see section 5 of this paper) truncate Fosse B's uppermost infill (Figure 6.6). These scours also cut through the infill of Fosse C.

In plan view, Fosse B presents an irregular ellipsoid shape with its major axis orientated subparallel to the strike of the outcropping Hythe Beds (Figure 6.4). Its northern edge is straighter than the southern one. Its northern edge actually extends along the southern slope of the ridge formed by the outcropping Hythe Beds in the seafloor, which also separates Fosse A from Fosse B (Figure 6.3 and Figure 6.4). That ridge is much thinner in the seafloor along the northern edge of Fosse B than further east or west. This seems to have been caused by a combination of the erosion produced by the incision of Fosses A and B with the folding and shortening induced by the reverse faulting, which reaches its maximum offset at the center of the Strait (See Chapter 4).

NE–SW cross-sectional profiles (i.e. parallel to the orientation of the Lobourg Channel) show general scoop-shape morphologies (Figure 6.3), with the maximum depth and dip (~20°SW) located along its northern slope. Note that all dip angles provided in this Chapter refer to average apparent dips measured in the seismic reflection data; they have been calculated by applying the average velocity used to build the Fosses Dangeard's isopach map.

Fosse B is carved into sub-horizontal strata, exhibiting neither faulting nor folding (Figure 6.6). Hence, both the incision and the shape of this palaeo-depression do not appear to have been geologically or structurally controlled. The infill of this palaeo-depression is interrupted by several internal erosions. We have identified 6 major erosional surfaces; i.e. from the basal erosional surface (Eb0) upward: Eb1–6 (Figure 6.3 and Figure 6.6). Only part of the sediments between the basal erosional surface (Eb0) and the first internal erosion (Eb1) remain in Fosse B. Internal erosional surface Eb1 truncates Eb0 in the southern half of Fosse B, incising directly into bedrock. Eb1 extends over the entire palaeo-depression, showing a sinusoid cross-sectional morphology. Interestingly, it incises more in the south (into bedrock) than in the north (into the Fosse's infill). This particular cross-sectional morphology is repeated by the younger internal erosional surface Eb4, which truncates all internal erosions underneath and incises into bedrock in the southernmost part of Fosse B (Figure 6.3).

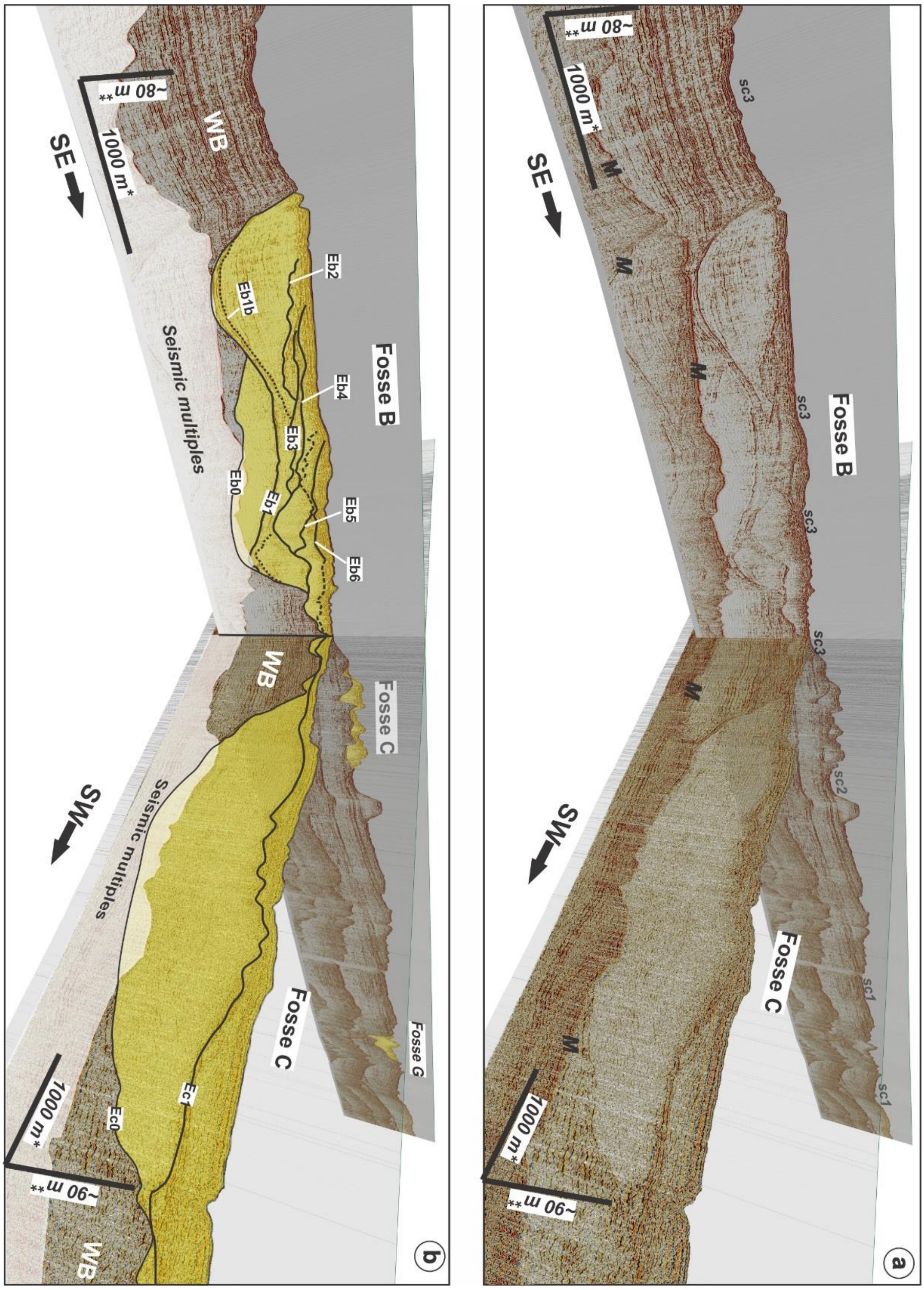

Figure 6.6. 3D plot of cross-cutting high-resolution single-channel seismic-reflection profiles 9 and 11 (see inset in Figure 6.3). Note that internal erosion surface Eb6 of Fosse B correlates with Ec1 in Fosse C, and the absence of other internal erosional surfaces in Fosse C. Yellow area: infills of Fosses B and C; Eb1, Eb2, etc.: internal erosional surfaces; WB: Wealden Beds; sc1, sc2 and sc3: longitudinal scours (see Figure 6.10 and Figure 6.14); M: seismic multiples. *Distance measured along different seismic profiles' orientations; **apparent differences in lengh and depth bars are due to 3D projection.

Internal erosional surface Eb6 also appears to have eroded large parts of Fosse B's older infill. Its spatial distribution and morphology is nevertheless unclear due to the bad quality of the seismic data near the surface. This erosion is however better defined in Fosse C, where it extends over the entire palaeo-depression as a relatively flat surface. Importantly, Eb6 is the only internal erosion that can be confidently correlated between Fosses B and C (Figure 6.6). The geometries of internal erosional surfaces Eb2, Eb3 and Eb5 are less well defined in the seismic reflection data, since significant parts of them have been truncated by younger internal erosions.

The seismic facies separated by the various internal erosional surfaces can be classified as follow:

- Eb0–Eb1: almost transparent seismic facies with no distinct reflections;
- Eb1–Eb2: high-amplitude reflections at the bottom that change to lower-amplitude well-defined parallel ones from horizon Eb1b upward. The latter marks a change of seismic facies within this unit, as it does not seem to truncate the reflections underneath;
- Eb2–Eb3, Eb3–Eb4 and Eb6–seafloor: discontinuous moderate to high-amplitude reflections;
- Eb4–Eb5 and Eb5–Eb6: diffused reflection with some localized discontinuous reflections.

The similarities between erosional surfaces Eb1 (Fosse B) and Ea1 (Fosse A) and the characteristics of the seismic facies above and below them, suggest that these two surfaces may have been incised by similar erosional events. In the same way, internal erosions Eb2, Eb3, Eb4 and Eb6 present similarities with Ea2, Ea3, Ea4 and Ea5 respectively. However, with the data at hand it is impossible to confidently correlate the different surfaces across Fosses A and B.

6.4.3 Fosse C

Fosse C is an irregular, 90–100 m deep, infilled palaeo-depression elongated across the strike of the outcropping bedrock (Figure 6.4b). This depression is incised entirely within the Lobourg Channel. Fosse C, as Fosses B and F, is carved mostly into the sub-horizontally stratified Weald

Clay unit of the Wealden Group (Figure 6.3 and Figure 6.6).

In plan view, Fosse C is elongated in the general orientation of the Lobourg Channel (Figure 6.4). As in Fosse B, its northern edge is much straighter than the southern one, extending along the outcropping Lower Greensand Formation. Cross-sections along its major axis (i.e. NE–SW) reveal scoop-shape morphology (Figure 6.3). In this case, however, the northern slope is less steep (dip: ~7°SW) than in Fosse B, although it is still ~2° steeper than its southern slope.

In contrast to Fosses A and B, only one internal erosion (Ec1) is clearly distinct in the infill of Fosse C. As mentioned above, Ec1 correlates with the youngest erosional surface observed in Fosse B, i.e. Eb6 (Figure 6.6). The rest of Fosse C's infill is rather homogeneous, consisting of diffused seismic facies, exhibiting a few low-amplitude, discontinuous reflections.

Fosses C and F cut across two WNW-trending fault systems deforming the bedrock underneath (Figure 6.4b). These faults seems to have had little influence on the incision and morphologies of these palaeo-depressions. Note that both palaeo-depressions are elongated across the faults strikes. In addition, connections between these palaeo-depressions occur through unfaulted areas, rather than along the faults' planes.

6.4.4 Fosse D

Fosse D is an 80–90 m deep, sediment-filled depression, ellipsoidal in shape in map view, and elongated along the strike of the outcropping bedrock (Figure 6.4). It connects with Fosse E by a narrow SW–NE corridor that shallows eastwards. This palaeo-depression is entirely incised outside the Lobourg Channel, demonstrating its independence from the erosional event(s) that carved the latter.

NNE–SSW cross sections reveal scoop-shape morphologies (Figure 6.3), with the northern slope (dip: ~30°SW) much steeper than the southern one (dip: 10–20°NE). NE–SW cross sections, on the other hand, exhibit almost semi-circular morphologies (Figure 6.7). The seismic reflection profiles also show that this palaeo-depression cuts through sub-horizontal stratified Hythe Beds. No tectonic structure traverses this palaeo-depression. The concentration of the erosion in this location and the shape of the incision are therefore unrelated to the bedrock geology and local tectonic structures.

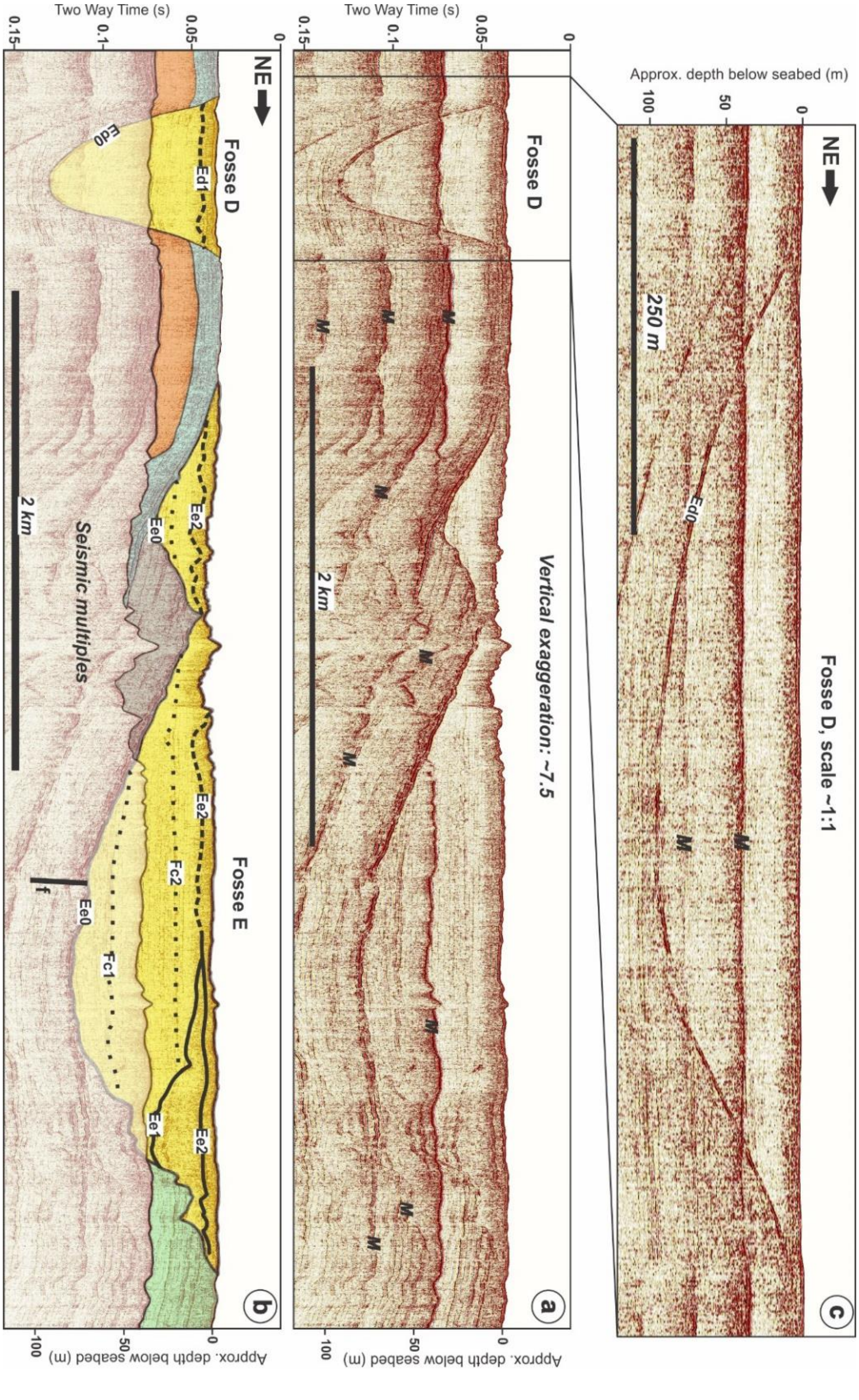

Figure 6.7. Non-interpreted (a) and interpreted (b) high-resolution single-channel seismic-reflection profile acquired across Fosses D and E (profile number 2 in Figure 6.3). c) Zoom on Fosse D. Note that the bedrock units underneath Fosse D are sub-horizontally stratified. Note also the absence of bedrock faulting below Fosse D. Black lines within yellow area: internal erosional surfaces; Fc1 and Fc2: seismic horizons marking possible changes of seismic facies. See Figure 6.4 and Figure 6.3 for other labels and meaning of colored areas.

Seismic facies infilling Fosse D are characterized by subparallel low-amplitude reflections. Only one possible internal erosional surface is observed in the seismic data just underneath the seafloor (named Ed1 in Figure 6.7). However, the geometry of this and the seismic facies above it are unclear due to the bad quality of the seismic reflection data near the seafloor.

6.4.5 Fosse E

Fosse E is a 1.5–4.5 km wide, 90–100 m deep sediment-filled depression elongated along the strike of the outcropping bedrock. This palaeo-depression is also carved outside the Lobourg Channel (Figure 6.4). NE–SW cross sections show that Fosse E comprises a series of connected scoop-shape buried depressions that shallow southwestward (Figure 6.3 and Figure 6.7). As the palaeo-depressions described above, its northern slope (dip: ~7°SW) is steeper than the southern one (dip: ~3°NE). Fosse E is shallower across the Lower Greensand units 'Sandgate Beds' and 'Hythe Beds' than elsewhere. These units, especially the Hythe Beds, are two of the bedrock units more resistant to erosion outcropping in the Strait, attesting to some degree of geological control of Fosse E's morphology. This is further evidenced by the dip of its southwestern slope, which seems to be controlled by that of the incised bedrock units (Figure 6.7). Its northeastern edge, on the other hand, incises into sub-horizontally stratified chalk. Bedrock faulting below this depression is much localized, suggesting little, if any, structural control of the erosion.

The infill of this palaeo-depression comprises two changes of seismic facies, labelled Fc1 and Fc2 (Figure 6.3 and Figure 6.7), and two well-defined internal erosional surfaces, named Ee1 and Ee2. Fc1 and Fc2 separate three distinct seismic facies; i.e. high-amplitude reflections between the basal erosion (Ee0) and Fc1, discontinuous to chaotic sub-horizontal reflections between Fc1 and Fc2, and low-amplitude sub-horizontal reflections above Fc2. These surfaces are less sharp than internal erosions Ee1 and Ee2 and do not seem to truncate the reflections lying underneath. Hence, we interpret them as changes in the sedimentary settings rather than significant erosional events. On the other hand, Ee1 and Ee2 clearly truncate the infill underneath (Figure 6.7). Ee1 presents scoop-shape cross-sectional morphologies and seems to be infilled by cross-bedded reflections. Ee2 exhibits flatter morphologies and extends over lager areas than Ee1 within Fosse E. The seismic facies above Ee2 are characterized by diffused reflections (Figure 6.7).

6.4.6 Fosse F

Fosse F is an asymmetrical, 120–140 m deep, sediment-filled depression presenting similar plan-view morphology to Fosse C (Figure 6.4b). It is mainly carved into Weald Clay, and half of its extent is located within the Lobourg Channel. Note that palaeo-channel ch1, which is associated with the last phase of valley incision along the Lobourg Channel (see section 6.5 of this chapter), truncates the uppermost infill of Fosse F (Figure 6.8).

The northeastern edge of Fosse F extends parallel to the strike of the outcropping Hythe Beds (Figure 6.4). Its western extremity trends towards Fosse D–E, suggesting that these palaeo-depressions might be connected to each other. Nevertheless, no data is currently available to corroborate that. NE–SW cross sections (i.e. across its major axis) show scoop-shape morphologies with the northeastern slope much steeper than the southwestern one (~17°SW vs. ~9°NE). The basal erosional surface also exhibits a series of steps, troughs and highs (Figure 6.8).

The seismic facies infilling Fosse F are quite homogeneous. Two internal erosions are however observed near the seafloor surface. These are, a minor channel-like, cup-shape erosional surface located at the northeastern slope of Fosse F, and a sub-horizontal, ~10 m deep, relatively flat erosional surface (Ef1) that is more widespread (Figure 6.8). Both the seismic facies above Ef1 and its morphology look similar to internal erosions Ee2 in Fosse E and Ed1 in Fosse D, suggesting that there may be a correlation between them (compare Figure 6.3, Figure 6.7 and Figure 6.8).

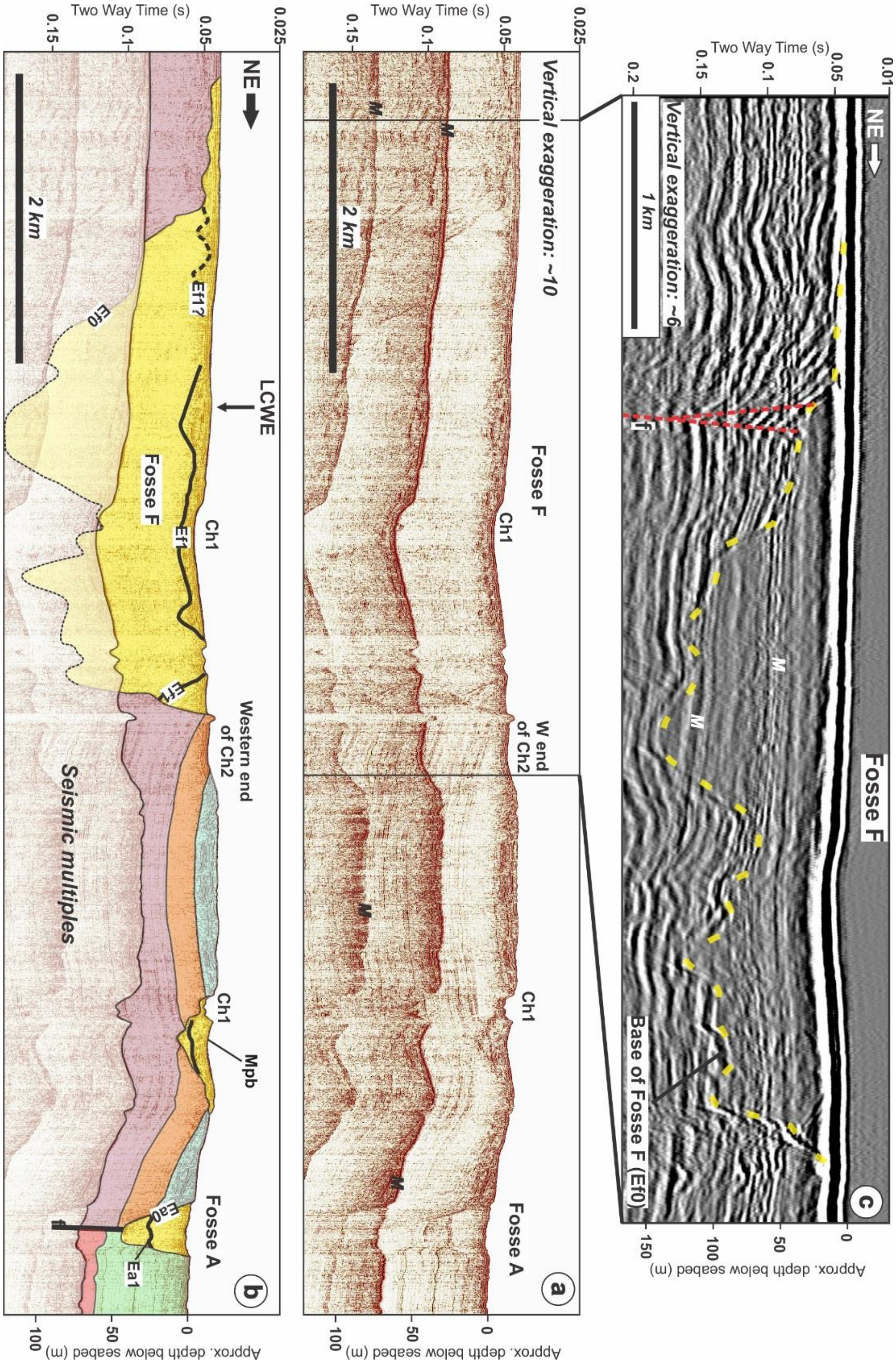

Figure 6.8. Non-interpreted (a) and interpreted (b) high-resolution single-channel seismic-reflection profile acquired across Fosse F and Fosse A (profile number 7 in Figure 6.3). c) Zoom on Fosse F extracted from a multi-channel deeper-penetration seismic-reflection profile acquired parallel to (~1 km to the north of) single-channel seismic-reflection profile 7. Ef1 and Ea1: internal erosional surfaces. See Figure 6.4 and Figure 6.3 for other labels and meaning of colored areas.

6.4.7 Fosse G

Fosse G is a NE–SW-oriented, elongated, 50 m deep, ellipsoidal (in plan view) buried depression located along the eastern slope of the Lobourg Channel and entirely carved into the Wealden Beds (Figure 6.4). This palaeo-depression is characterized by a semicircular to subtle scoop-shape morphology in cross section, with the northeastern slope slightly steeper than the southwestern one (Figure 6.9).

The infill of Fosse G exhibits two distinct seismic facies. That is, a relatively thin package of high-amplitude horizontal reflections lying atop an acoustically almost transparent unit. However, the quality of the seismic data does not allow determining whether the surface separating them represents an internal erosional surface or merely a facies transition; i.e. a change in the nature of the sediments.

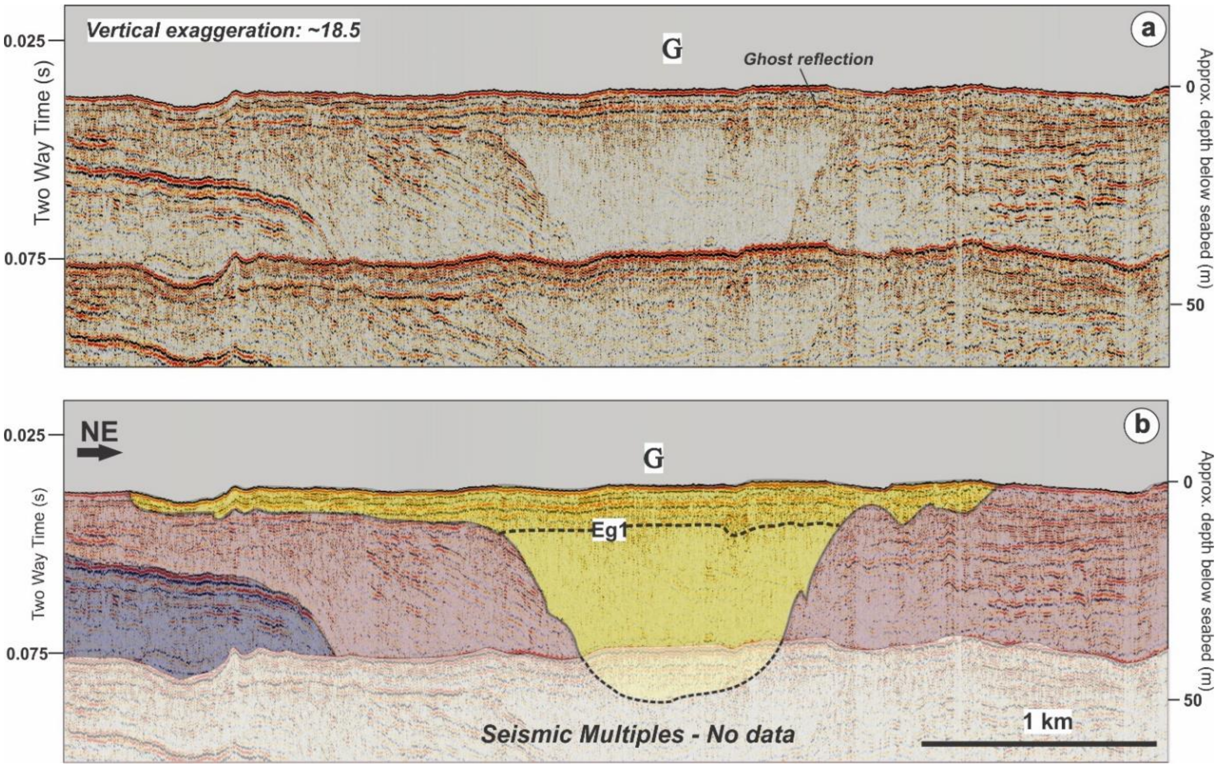

Figure 6.9. Non-interpreted (a) and interpreted (b) high-resolution single-channel seismic-reflection profile collected along Fosse G (profile number 12 in Figure 6.3). Eg1: internal erosional surface; ghost reflection: seafloor repetition near the surface (artefact). See Figure 6.4 and Figure 6.3 for meaning of colored areas.

6.5 The Lobourg Channel

Despite the fact that some of the Channel palaeovalleys have been partially or totally infilled with sediments (e.g. James et al., 2002; Mellett et al., 2013), it is possible to reconstruct their geomorphology by using bathymetric data (Figure 6.1; see Mellett et al., 2013; Collier et al., 2015). This is especially true in the Dover Strait, where Holocene sediments are scarce and older Quaternary ones are mostly concentrated in the Fosses Dangeard (Figure 6.4). In this section, we describe in detail the geomorphology of the Lobourg Channel. To ease the descriptions, we have divided the Dover Strait in three parts; i.e. northern, central and southern Dover Strait (Figure 6.10a).

6.5.1 Northern Dover Strait

In the northern Dover Strait, the Lobourg Channel consists of a ~25 km wide, NE–SW- oriented palaeovalley, comprising three sharp streamlined escarpments (E1, E2 and E3) and some NE–SW-oriented streamlined islands carved into chalk (Figure 6.2 and Figure 6.10).

On the one hand, escarpments ‘E1’, which forms the eastern edge of the Lobourg Channel, and ‘E2’ are 10–15 m high and exhibit NE–SW orientations (Figure 6.10 and Figure 6.11). On the other hand, escarpments ‘E3’ and the ‘Lobourg Channel western edge’ (LCWE) are, respectively, 5–10 m and 15–25 m high. They trend NNE–SSW and run southward obliquely to E1 and E2. In fact, the LCWE and E3 escarpments delimit a prominent ~3.5 km wide, box-shape palaeovalley carved into bedrock (Figure 6.10 and Figure 6.11), the eastern edge of which (E3) seems to truncate E1 and E2 in the central Dover Strait. In this study, we refer to that palaeovalley as “Lobourg inner channel” (Lic).

The northwestern extents and morphologies of E1 and E2 (especially that of E1) are poorly constrained, since sandbanks and dunes of Holocene age cover large parts of their extents (Figure 6.10). E3 and LCWE are, however, sharply marked in the seafloor, displaying rather rectilinear morphologies. Topographic cross-sections across the four main escarpments (LCWE, E1, E2 and E3) show morphologies similar to strath-terraced fluvial valleys (e.g. Hancock and Anderson, 2002), with all the terraces located on the eastern side of the valley (Figure 6.11). Importantly, these terraces are not formed by gravels or sand aggradations; they are carved into bedrock. In this study, we have named the terrace-like platform delimited by escarpments E1 and E2 as LC1, while the one defined by E2 and E3 is referred to as LC2.

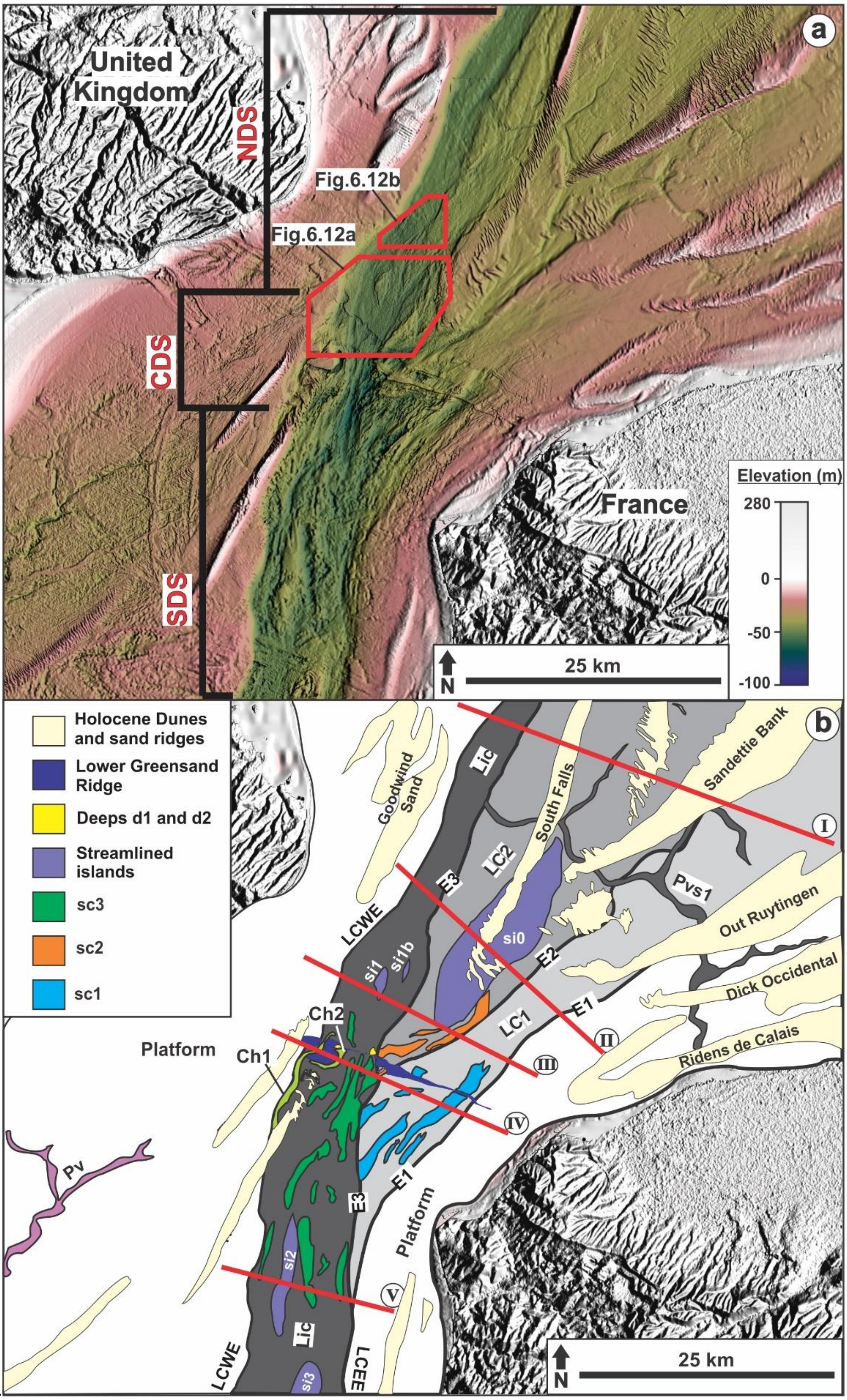
a
United Kingdom
NDS
CDS
SDS
Fig.6.12b
Fig.6.12a
France
Elevation (m)
280
0
-50
-100
25 km
N
b
Holocene Dunes and sand ridges
Lower Greensand Ridge
Deeps d1 and d2
Streamlined islands
sc3
sc2
sc1
Goodwind Sand
Lic
South Falls
Sandettie Bank
LC2
E3
Pvs1
Out Ruytingen
LCWE
si0
si1
si1b
E2
E1
Dick Occidental
LC1
Ch2
Ch1
Platform
Ridens de Calais
I
II
III
IV
V
Pv
Platform
si2
si3
LCEE
25 km
N

Figure 6.10. Geomorphological interpretation of the Lobourg Channel. a) Merged bathymetry of the Dover Strait area gridded at 40 m cell size. b) Geomorphological interpretation. Note that escarpment E3 truncates escarpments E1 and E2. With the exception of si0 and si1b, which were not previously identified, streamlined islands (si) are labelled after Collier et al. (2015). Major tidal sand ridges are labelled after British Admiralty Nautical Charts. Red lines in (b) indicate the location of topographic profiles shown in Figure 6.11. Pvs1: palaeovalley system discharging into the Lic; Pv: valley network incised into platform described in Chapter 5; E1, E2 and E3: Escarpments formed during different phases of valley incision along the Lobourg Channel; sc1, sc2 and sc3: longitudinal scours associated with valley incisions LC1, LC2 and the Lic; LCWE: Lobourg Channel Western Edge; LCEE: Lobourg Channel Eastern Edge; Red lines in (a): areas shown in Figure 6.12; black lines in (a): extents of the northern Dover Strait (NDS), central Dover Strait (CDS) and southern Dover Strait (SDS).

Both LC1 and LC2 are traversed by a minor palaeovalley system (i.e. Pvs1), which cuts through E1 and E2 perpendicularly to their main orientations (Figure 6.10). This system is partially covered by sandbanks and dunes, and it appears to have been a tributary of the Lic. Streamlined islands are only observed on LC2 and within the Lic (Figure 6.10). These include a ~15 km long, 5 km wide, major streamlined island (si0), and some ~2.5 km long, ~1 km wide, minor ones (e.g. si1 and si1b). All streamlined islands identified in this area present similar morphologies to those described by Collier et al. (2015) carved into chalk; i.e. tear-drop shapes, with their major axis oriented in the direction of the inferred palaeo-flow. On the one hand, si0 is located within LC2 and its major axis is subparallel to escarpment E2, suggesting that the formation of si0 is associated with the processes that also carved E2. On the other hand, s1 and s1b are located within the Lic, which has similar orientation to their major axis. This indicates that the incision of s1 and s1b was most likely associated with erosional processes channeled through the Lic. si0 exhibits a series of ENE–WSW-oriented scours carved along its southeastern edge and front part. These scours end in the west, at their intersection with E3, in a 20-m-deep depression (d1) incised into Fosse A's infill (Figure 6.14).

The overall morphology of the Lobourg Channel suggest the occurrence of three major phases of valley incision. That is, LC1 and LC2 appear to represent remnants of two palaeovalleys (or of two phases of valley incision) that were truncated by the almost-straight, NNE–SSW-oriented valley defined by the Lic (Figure 6.10). The original width and morphology of palaeovalleys LC1 and LC2 are unknown, since no trace of their western slopes remains. Nonetheless, the geomorphology of the Lobourg Channel suggest that, at least, LC2 may have been rather wide. Indeed, if its western slope coincided with the present-day LCWE, its width could have reached up to ~20 km in the northern Dover Strait (Figure 6.10 and Figure 6.11).

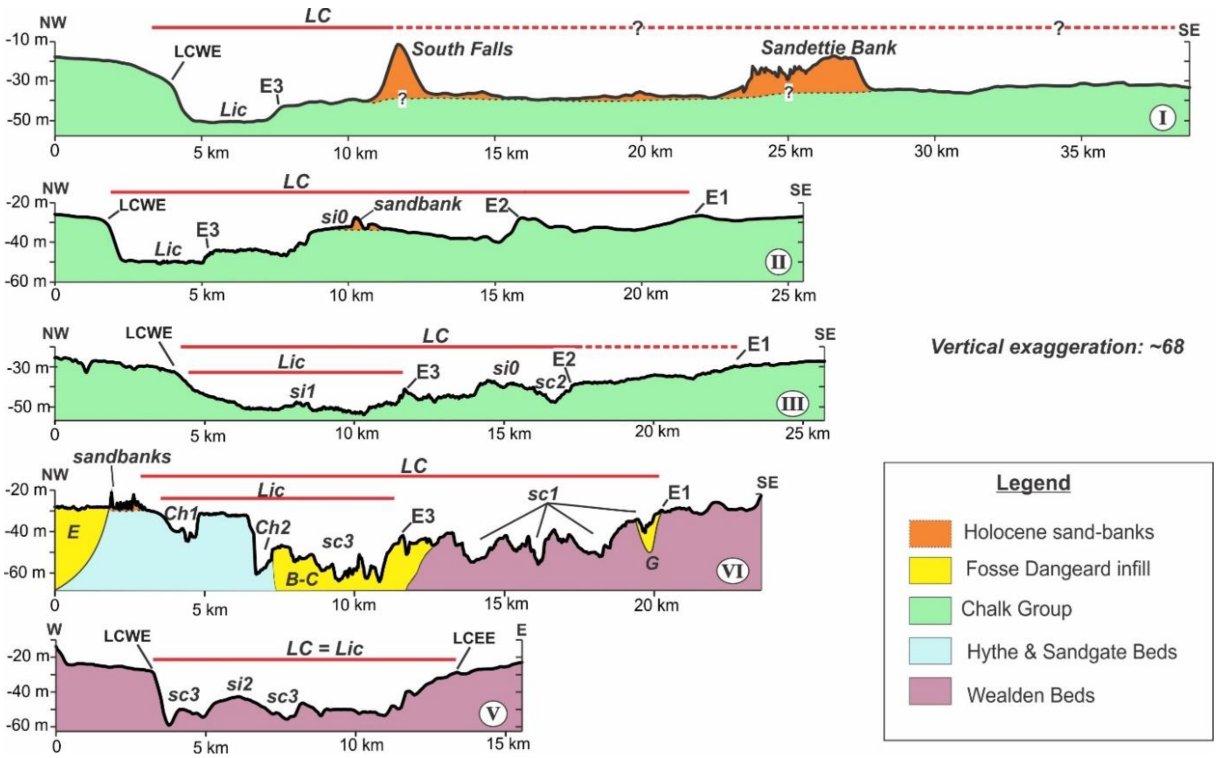

Figure 6.11. Geologically interpreted topographic profiles across the Lobourg Channel (see Figure 6.10b). E, B-C and G: Fosses E, B and C; LC: interpreted width of the Lobourg Channel; Lic: Lobourg inner channel. See Figure 6.10 for other labels.

The orientation of E1 and E2 suggests that palaeovalleys LC1 and LC2 might have narrowed southwestward as they approached the central Dover Strait (Figure 6.10). Contrarily, the width of the Lic is rather constant along the northern Dover Strait. Topographic cross-sections across it reveal that the Lic is characterized by quasi-symmetrical slopes (dips: ~1.7°) and a flat bottom (Figure 6.11). It widen significantly though at the southern end of si0, passing from ~3.5 km to ~7.5 km wide. Importantly, the Lic extends over tens of kilometers across the southern North Sea and northern Dover Strait, exhibiting similar relief and shape across Paleogene and Neogene clays and Cretaceous chalk (Figure 6.2 and Figure 6.10).

Other remarkable features observed in the northern Dover Strait are several sets of parallel streamlined ridge-and-groove linear bed-forms (Figure 6.12 and Figure 6.13). These features are oblique to ship-track linear artefacts, dunes and other Holocene sedimentary bodies. This indicates that they are neither artefacts nor Holocene sedimentary features. Considering that most river systems running into the southern North Sea possibly entered the English Channel through the Dover Strait during the Last Glacial Maximum (Gibbard et al., 1995), it is unlikely that these features are remnants of sand ridges formed during the Eemian marine highstand. Hence, our preferred interpretation is that these features represent bedrock-eroded longitudinal grooves. However, they are not traversed by any of the seismic reflection profiles available for this study, preventing us to confirm that interpretation. The majority of these features are located within the Lic. Some of them are however located outside it. Importantly, those situated outside the Lic are truncated by its western edge, suggesting that these features formed before the incision of the Lic (Figure 6.12).

Individual ridges and grooves are generally up to 30 m wide, with amplitudes ranging between 0.5 and 1.5 m (Figure 6.13). Their lengths are however unknown, as sand dunes and other younger sedimentary bodies cover large parts of them (Figure 6.12). They have a consistent ENE alignment and extend subparallel to the axis of the Lic. Their azimuths vary slightly from one set to another: Azimuths of set 'La' (incised outside the Lobourg Channel) range between N32°E and N36°E, those of set 'Lb' are between N40°E and N45°E, and those of set 'Lc' are between N37°E and N41°E.

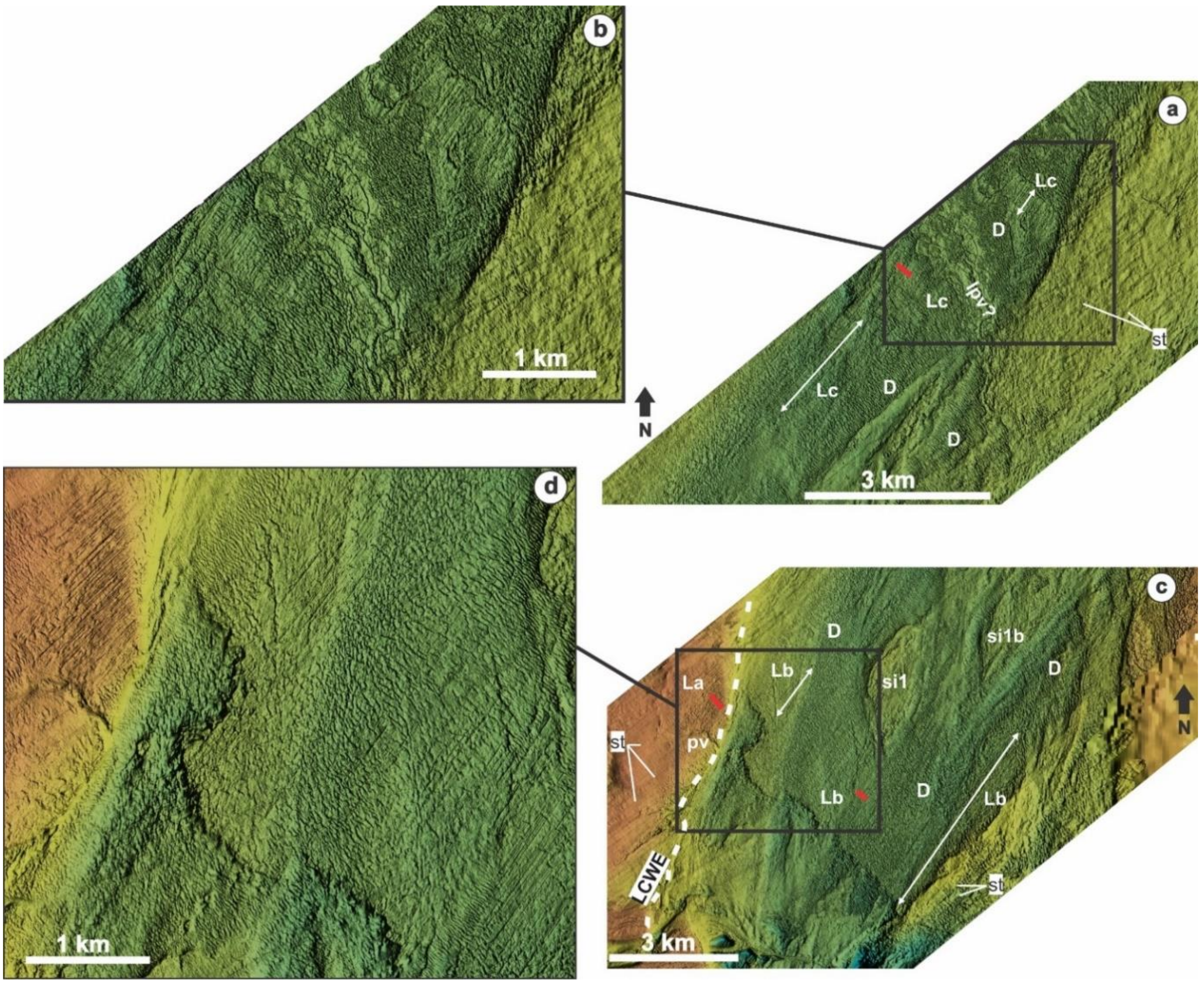

Figure 6.12. Bathymetric data gridded at 1.5 m showing linear groove-and-ridge bed-forms (i.e. La, Lb and Lc) carved into chalk. Orientation of lineations is indicated by double-headed arrows. Note the apparent truncation of the lineation set "La" by the western edge of the Lic (LCWE). Note also that the groove-and-ridge bed-forms are covered by, and have different orientation than, dunes (D) and minor infilled palaeovalleys (e.g. Ipv). They present different orientations from linear artefacts (st) as well. Red lines: topographic profiles plotted in Figure 6.13.

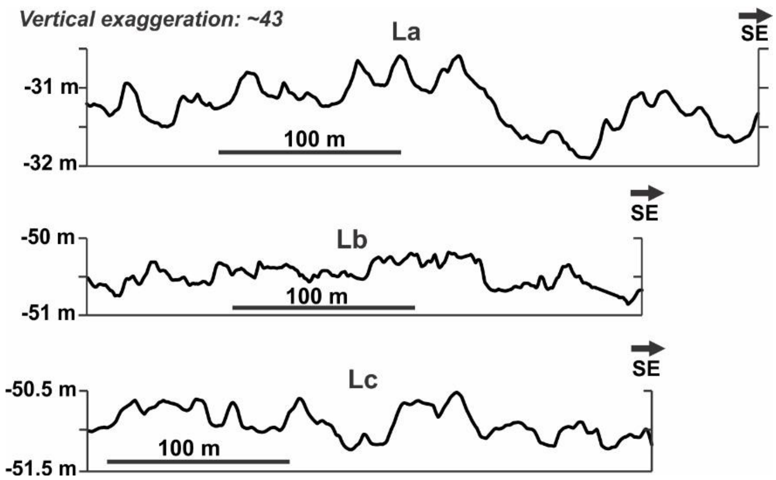

Figure 6.13. Topographic cross sections across the different sets of groove-and-ridge bed-forms. See Figure 6.12 for location.

6.5.2 Central Dover Strait

Contrasting with the rather homogeneous bedrock outcropping in the northern (i.e. chalk) and southern (mudstones and clays) Dover Strait, the bedrock outcropping in the central Dover Strait is characterized by a range of sedimentary formations (Figure 6.4), exhibiting different resistance to erosion. In addition, the Fosses Dangeard are located in that area.

The central Dover Strait coincides with the northern Lower Greensand limb of the Weald-Artois anticline (Figure 6.2), and it is traversed by the main tectonic structures composing the submarine North Artois Shear Zone. It also marks a major change in the orientation of the Lobourg Channel (Figure 6.10). The Lic and escarpment E1 bend counterclockwise at their path through the outcropping Lower Greensand formation, continuing southwards with nearly north–south orientations. The overall geomorphology of the Lobourg channel also experience a major change from the center of the Strait southward. For instance, escarpments E2 and E1 are truncated in the central Dover Strait by the Lic, which defines the entire Lobourg Channel in the southern Dover Strait (Figure 6.10).

The erosional features carved within the Lobourg Channel in the central Dover Strait are also significantly different from those carved within this palaeovalley further north (see Figure

6.10, Figure 6.12 and Figure 6.14). In particular, several of the stratigraphic units outcropping in that area form prominent scarps, ridges and grooves, extending perpendicularly to the orientation of the Lobourg Channel and attesting to differences in the resistance to erosion of the various bedrock units. Further, neither the ridge-and-groove linear features nor the tear-shape streamlined islands are found in the center of the Strait.

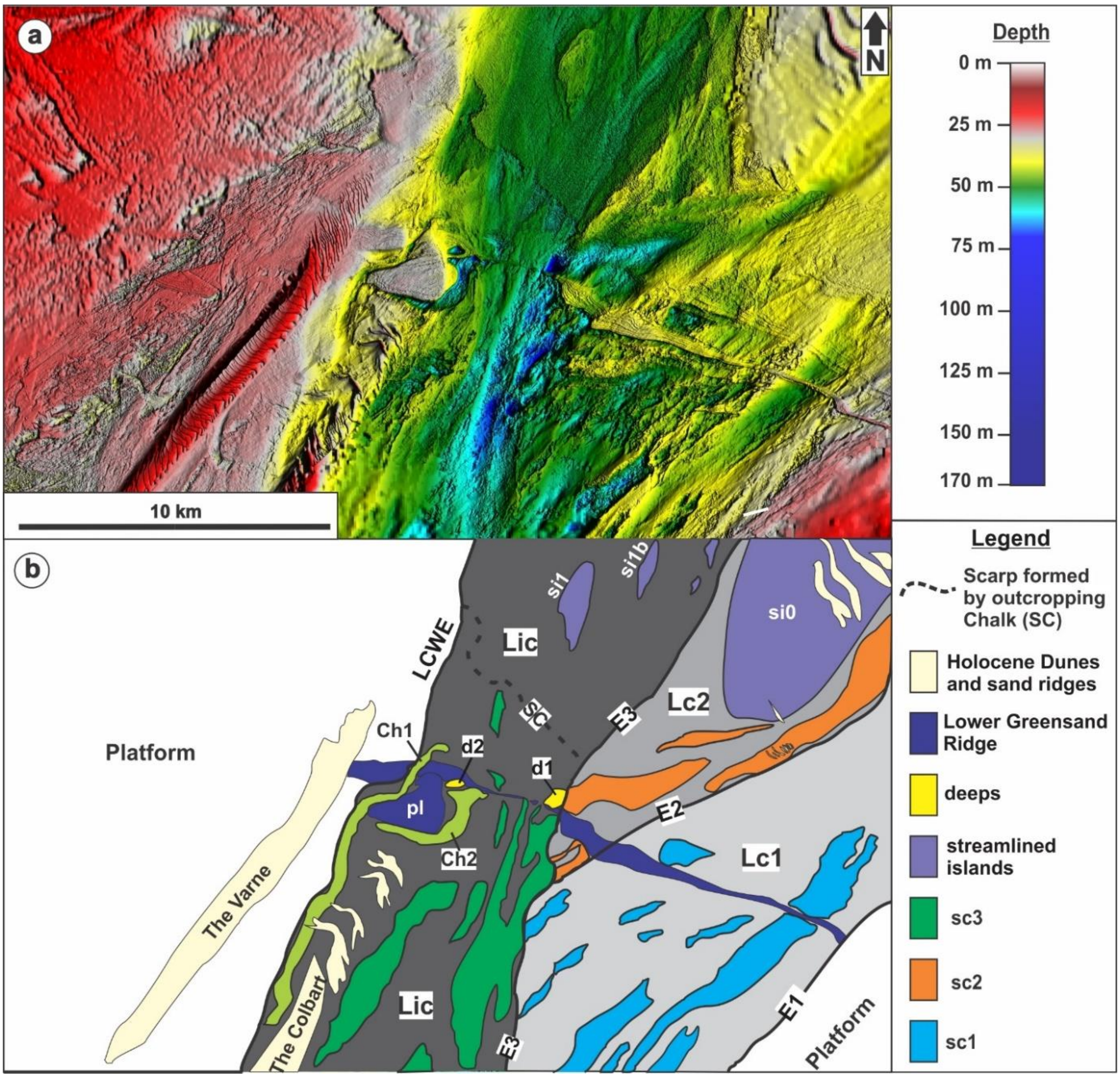

Figure 6.14. a) Bathymetric merged of the central Dover Strait (i.e. same area than the one plotted in Figure 6.4; several resolutions). b) Geomorphologic interpretation. CG: Scarp formed by outcropping Chalk Group units; Ch1 and Ch2: inner channels. For other labels, see Figure 6.10.

The most significant erosional features located in the central Dover Strait are carved within the Lic (Figure 6.14). They include a series of N–S-oriented longitudinal scours (sc3), two deeply incised inner palaeo-channels (Ch1 and Ch2) and a 20 m deep ovoid depression (d2).

Scours sc3 are kilometer-scale elongated incisions, exhibiting depths at the seafloor of up to 10 m. These features cut through Gault Clay, Lower Greensand, the uppermost infills of Fosses A, B and C, and the Wealden Beds (Figure 6.4 and Figure 6.14). Ch1 and Ch2 are 0.5–1 km wide, amphitheater-head palaeo-channels incised along the western half of the Lic. Ch1 is characterized by a linear to slightly sinuous morphology and relief of 10–15 m at the seafloor. This channel cuts through Fosse F's upper infill and the Cretaceous Lower Greensand and Wealden Beds Formations (Figure 6.4 and Figure 6.14). It extends over, at least, 9 km along the western edge of the Lobourg Channel. However, its total extent is unknown, as large tidal sand ridges cover its southern part. Ch2, on the other hand, is restricted to the eastern and southern part of the plateau formed between Ch1 and Ch2 by the outcropping Hythe Beds (Figure 6.14). This channel is mostly carved into Atherfield Clay and Wealden Beds (Figure 6.4). It extends over ~4 km and shows ~20 m of relief at the seabed (Figure 6.11). Ch2 branches to the northeast into two amphitheater-shaped heads, the headwall of which reaches maximum height of ~24 m and has slopes of 10°–15° (see Chapter 5). Immediately northward of its headwall, the ~600-m-wide ovoid depression "d2" is eroded ~20 m into Hythe Beds bedrock (Figure 6.14).

Other significant erosional features carved in the central Dover Strait area are longitudinal scours sc1, which are located within LC1 and present similar orientation to E1. These features, as LC1, are truncated by the Lic, suggesting that their formation is related to the valley incision that shaped LC1. Note that these scours truncate the infill of Fosse G (Figure 6.11), indicating that their formations, and so the first phase (LC1) of valley incision identified in the Lobourg Channel, is younger than the formation of the Fosses Dangeard and the sediments infilling Fosse G. However, the lack of sedimentary data prevents correlations between the infill of Fosse G and those of other Fosses. Hence, we have not been able to determine whether the valley incisions that carved LC1 and LC2 happened following the last infilling episode identified in Fosses A and B or, rather, they were contemporary to the incision of one or several of the internal erosional surfaces interrupting the infills of those Fosses.

Contrarily to previous erosions, the Lic cuts through all outcropping units, including the resistant Hythe Beds (Figure 6.4). In fact, the ridge formed by that stratigraphic unit across the width of the Strait is almost completely absent in the eastern half of the Lic (Figure 6.14). Even in the western half of the Lic, where the width of that ridge is at its maximum owing to local folding, Ch1 and the ovoid depression d2 incise, respectively, ~5 m and ~20 m into that unit, attesting to the extreme erosion that carved these features. The Lic thus represents the most intense erosional event currently observable in the seafloor that took place following the formation of the Fosses Dangeard.

6.5.3 Southern Dover Strait

The Lobourg Channel in the southern Dover Strait cuts through Wealden Beds and Kimmeridge Clay Formations, which have rather similar composition. In this area, the Lic occupies the entire Lobourg Channel (Figure 6.10). The Lobourg Channel passes thus from a 20-km-wide, terraced palaeovalley in the north and central Dover Strait to a 10-km-wide, single channel in the south. Topographic cross sections across the southern Lobourg Channel show irregular box-shape morphologies with steeper slopes in the west (3–4°) than in the east (1–2°).

The margins of the Lobourg Channel in the southern Dover Strait are up to ~30 m high (Figure 6.11). Erosional features within the Lobourg Channel in that area include linear and sinuous kilometer-scale scours and elongated streamlined islands (Figure 6.10). This change in the incision pattern is probably partly due to the lower resistance to erosion and more homogeneous composition of sedimentary formations composing the seafloor in the south (Figure 6.2; see Collier et al., 2015).

6.6 Discussion

The present-day geomorphology and recent Quaternary geology of the Dover Strait comprises a series of palaeovalleys and palaeo-depressions that truncate one another, attesting to a succession of several major erosional events. Here below, we discuss the origin of the various major erosional features described in this study based on combinations of our observations with the Quaternary geological/palaeogeographic context stablished by previous works performed in this area.

6.6.1 Formation and infilling of the Fosses Dangeard

The tight seismic grid analyzed in this study has permitted accurate 3-dimensional morphological mapping of the Fosses Dangeard. This detailed analysis has revealed, for instance, that Fosses A, B and F are 20–40 m deeper than previously thought (compared with

Destombes et al., 1975; Smith, 1985). The high-resolution seismic-reflection data has also allowed seismic facies characterization of the infills of the various palaeo-depressions. This has led to the establishment, for the first time, of the number, morphology and spatial distribution of the various erosional surfaces interrupting the infill of the Fosses Dangeard. In addition, the combination of our results with previous studies lead us to corroborate or reject a series of previously postulated hypotheses.

As we already mentioned in Chapter 5, the remarkable depth of the Fosses Dangeard rules out erosion by tidal and/or marine currents as its cause. We can also definitively exclude a tectonic origin, since Quaternary tectonic deformations along the various identified tectonic structures appear to be negligible. Bedrock faults exist though below Fosses A, C, E and F (Figure 6.4). The presence of these structures may have induced a local increase of the erosion rate along their planes by differential erosion. Nonetheless, deep incisions happened indistinctly in zones with and without faults, precluding a direct relationship between the formation of the Fosses Dangeard and the faulting.

The morphology and isolation of the Fosses Dangeard shows no similarity with glacial tunnels or with glacial palaeovalleys (compare with features shown in Praeg, 2003; Lonergan et al., 2006; Kristensen et al., 2007). Moreover, the Fosses Dangeard are carved into bedrock, excluding glacial kettle holes. Hence, the combination of these observations with recent palaeogeographic reconstructions (e.g. Gibbard and Cohen, 2015) rules out glacial erosion too. The morphology and exceptional depth of the Fosses Dangeard also differentiate them from other isolated incisions of possible fluvial origin carved further southwest within the Channel palaeovalley system (see Mellett el al., 2013). Another particularity of the Fosses Dangeard is the fact that they cut through a range of sedimentary formations regardless their resistance to erosion (Figure 6.3). For instance, Fosse D is carved into sub-horizontally stratified, tectonically non-deformed limestone units.

The restriction of these palaeo-depressions to a belt extending along the barrier that was purported to have separated the English Channel from the southern North Sea until MIS 12 (e.g. Gibbard and Lewin 2016), and their significant depth/size support the model in which these palaeo-depressions were incised by large waterfalls. This model is reinforced by their particular morphologies, which strongly resemble plunge pools carved at the base of natural and modelled waterfalls formed due to the presence of major steps within rivers and high-magnitude flood flows (e.g. Alexandrowicz, 1994; Lamb et al., 2007; Lamb, 2008; Pagliara et al., 2008; Strasser et al., 2008; Baker, 2009). However, the depths of the Fosses Dangeard are rather unusual. Depths of other natural plunge pools associated with waterfalls induced by steps in flooded terrains and/or along large rivers range between a few meters to a few tens of meters (Alexandrowicz, 1994; Lamb et al., 2007; Lamb, 2008; Strasser et al., 2008; Baker, 2009). In those contexts, plunge-pool erosion is associated with vertical impact of falling water and sediments, plucking of fractured blocks and abrasion (Whipple et al., 2000). According to analogic models, the depth/size of the incision seems to be determined by the height of the waterfall, the volume of water falling, the sediment carried by the water and the composition of the substratum (i.e. Whipple et al., 2000; Lamb, 2008; Pagliara et al., 2008).

We propose that the unusual depth of the Fosses Dangeard might have resulted from the combination of the significant height of the waterfall, the large volumes of water falling, the relatively high resistance to erosion of the ridge's top (chalk), the possible high content of sediments in the water, and the extreme erosion induced by the megaflood that followed the waterfall phase. We discuss these possible factors here below.

The height of the Weald–Artois ridge across the Dover Strait is unknown. However, based on the height of the cliffs at either sides of the Strait, it has been proposed that it may have reached heights of ~30 m above present sea level (Gibbard and Cohen, 2015). Hence, adding the minimum depth (30–40 m) of the seafloor across the Dover Strait, it is likely that the waterfall was up to 60–70 m high. The relative height of the Weald-Artois ridge may also have been enhanced during glacial maximum due to glacial isostatic adjustments.

The volume of water cascading over the Weald–Artois ridge is unknown too, although it was certainly significant due to the continuous water input of northwestern European palaeo-rivers and episodic ice melting (Toucanne et al., 2009a; Toucanne et al., 2009b; Gibbard, 1995; Murton and Murton, 2012; Gibbard and Cohen, 2015).

The chalk that possibly formed the top of the waterfall and the less resistant and heterogeneous bedrock at its base might have also been determinant factors for the formation of such deep depressions. That is, the relative low erodibility of the chalk may have resulted in slow headward erosion of the waterfall wall, inducing longer exposures of its base, which was composed mostly of mudstones, sands and clays, to water and sediment impacts.

Finally, the erosional power of the waterfall was most likely increased at its base by swirling water charged with sediments and fragments of ice and rocks. These sediments came most likely from remobilization of sediments previously deposited at the base of the ridge and those composing some of the poorly consolidated subunits of the Wealden Beds and Lower Greensand Formations. In addition, the water may have contained flints disaggregated from the Chalk Group by dissolution, rocks fragments resulted from seasonal frost weathering of the ridge, ice fragments, etc. The volume of water and sediments would have increased by several order of magnitude during a megaflood event. Hence, if the breach of the Weald–Artois ridge induced a megaflood, this would have most likely caused intense erosion of the plunge pools, further deepening them.

We interpret the Fosses Dangeard as the result of plunge-pool erosion formed at the base of waterfalls that were retreating northeastward by the action of headward erosion. That is, cataracts retreating from Fosses C, D, F and G toward the northern edges of Fosse A and E, which most likely represent the last position of the waterfall's wall before the opening of the Dover Strait (Figure 6.15). We base this interpretation, firstly, on the fact that all Fosses exhibit steeper slopes in the north than in the south, which indicates that the palaeo-flow that incised them came from the north (see Lamb et al., 2007; Pagliara et al., 2008). Secondly, the northern edges of Fosses C, D, F and G, similarly to those of Fosses A and E, align with one another, suggesting that a linear barrier should have been located at some point along those edges. Finally yet importantly, it is very difficult to explain the location, depth and morphology of Fosse D without supposing that the cataract's wall was once located along Fosse D's northern edge. Indeed, lots of energy would be needed to form that palaeo-depression in such a hard, structureless substratum (see Pagliara et al., 2008).

The different morphologies of the palaeo-depressions are explained by the different resistance to erosion of the stratigraphic units into which they are carved. That is, waterfall headward erosion will be slow in resistant rocks, such as Hythe Beds and chalk, favoring the concentration of the erosion in a specific area over longer periods of time (see Alexandrowicz, 1994). This would result in deep, localized depressions (i.e. Fosses D and Fosse A). Whereas, in less resistant bedrock, such as the Wealden Beds or Gault Clay units, incisions will tend to be wider, deeper and elongated in the direction of the flow (e.g. Fosses B, C, E and F), since headward and plunge-pool erosions will be faster and incise more easily into those materials (see Alexandrowicz, 1994).

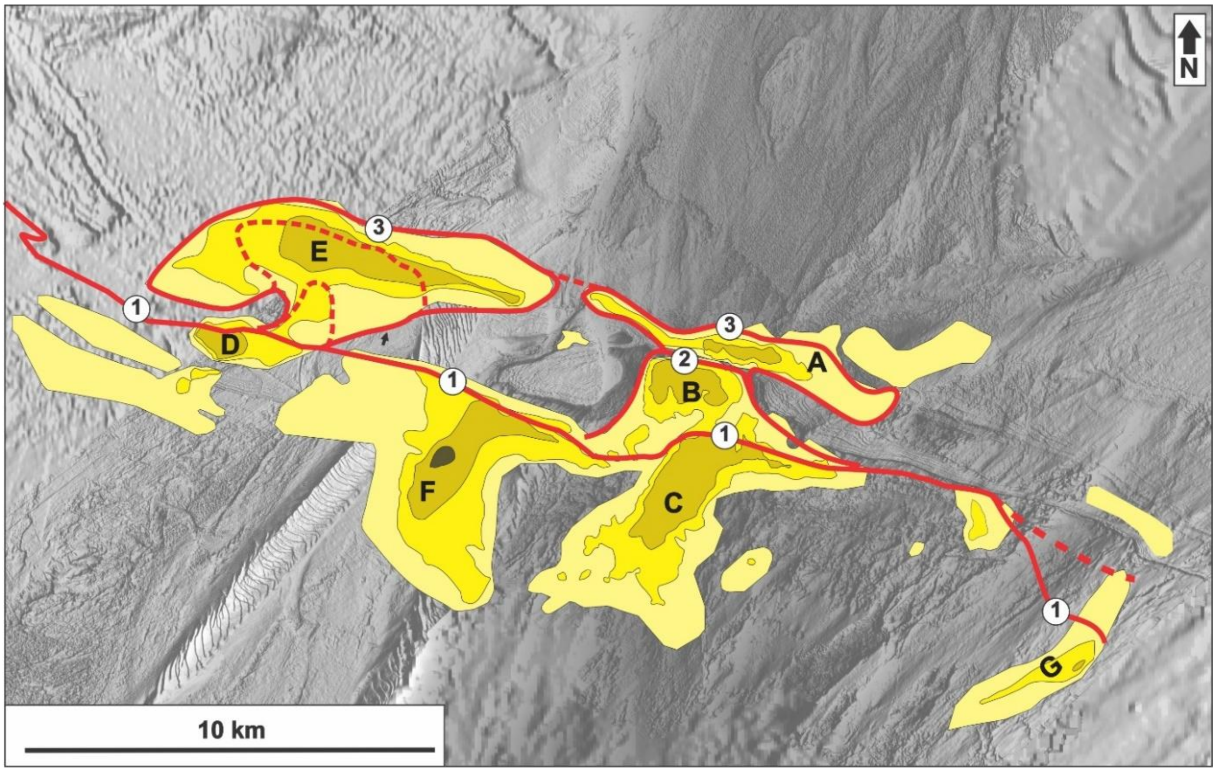

Figure 6.15. Hypothetical knickpoint headward erosion during the waterfall phase that incised the Fosses Dangeard. Red line 1 indicates the position of the waterfall's wall during the incision of Fosse D and last phases of incision of Fosses F and C. Red Line 2 shows the location of the waterfall during last phase of incision of Fosse B. Line 3 indicates the last position of the waterfalls (i.e. last phase of incision of Fosses A and E) before the breach of the Weald-Artois ridge and the initial opening of the Dover Strait.

The relatively high resistance to erosion of the Hythe Beds seems to have also prevented the northward extension of Fosse F and, thus, its connection with Fosse A. Indeed, that stratigraphic unit exhibits the widest outcrop observed in the Dover Strait along Fosse F's northern edge due to local folding and faulting (Figure 6.4). Contrarily, the combination of folding and reverse faulting has significantly narrowed the Lower Greensand and Gault Clay formations immediately to the east of Fosse F. This local bedrock configuration would therefore have favored a further northward progression of the incision in that area, thus explaining the continuation of Fosse C into Fosse B, and the almost connection of the latter with Fosse A.

Once passed the more resistant Hythe Beds, the northward retreating cataracts would have fallen right into a faulted zone in the center of the Strait and into the relatively less resistant Sandgate Beds, Folkestone Beds and Gault Clay units in the west. This would have favored deep incision again, resulting in the formation of Fosses A and E. In the case of Fosse A, the proximity of the Chalk Group to the Hythe Beds along the fault plane would have limited the northward progression of the incision, thus constraining the width of Fosse A. Fosse E, on the other hand, extends northward across gently north-dipping Hythe Beds, Sandgate Beds, Folkestone Beds and Gault Clay units. That is, the resistance to erosion of the bedrock across this palaeo-depression increases from south to north, thus explaining the northward widening and deepening of this incision (Figure 6.4).

The waterfall phase appears to have ended at some point during the incision of Fosses A and E. Importantly, the northernmost edges of these palaeo-depressions are aligned to each other and subparallel to the strike of the Chalk Group. This suggests that the waterfall phase ended at the same time in all locations of the ridge. Otherwise, the erosion would have continue incising further northward in the locations where the water continued falling. This is consistent with an abrupt breach of the Weald–Artois ridge and the generation of a megaflood flow in the eastern English Channel. Once the ridge was opened, water would have flowed almost exclusively through the breach (i.e. the lowest point across the ridge). We propose thus that the initial breach happened in the center of the Strait, since the concentration of subsequent major scouring-and-infilling episodes in Fosses A and B suggests that significant fluvial systems were concentrated in that area following their initial incision (see below).

The analysis of the seismic units infilling the Fosses Dangeard suggest the occurrence of 6 major erosional-and-infilling episodes in Fosses A and B, at least 2 in Fosse E, and a minimum of one in the remainder of the Fosses. It is thus evident that Fosses A and B were not subjected to similar infilling-and-scouring sequence than the rest. Two of the erosional surfaces (i.e. Eb1 and Eb4) interrupting the infills of these Fosses were rather intense, incising both infill and bedrock. Contrarily to the basal erosional surface (Eb0) of Fosse B, Eb1 and Eb4 incise deeper downstream than upstream (Figure 6.3). Note that topographic cross sections across the headwall of palaeo-channel Ch2, which was most likely incised by northward retreating waterfalls (see Chapter 5), also exhibit opposite morphology to Eb1 and Eb4. This suggests that erosional surfaces Eb1 and Eb4 were not incised by waterfalls. Instead, their cross-sectional morphology significantly resembles scours incised by high-magnitude flood erosion and/or by a sudden increase of flow velocity and turbulence along a river (see Eilertsen and Hansen, 2008). Based on that, we interpret Eb1 and Eb4 as scours incised by high-magnitude flood flows or high-velocity (highly erosive) fluvial flows. This evidences the occurrence of at least two episodes of intense fluvial erosion and/or high-magnitude flood erosion in the center of the Strait in the span between the formation of the Fosses Dangeard and the incision of the Lic. More importantly, the fact that none of the internal erosional surfaces identified in any of the Fosses Dangeard resemble any of their basal erosional surface suggest the occurrence of only one phase of waterfall. This is consistent with palaeogeographic reconstructions, which propose that the proglacial lake inundating the southern North Sea during the Saalian glacial maximum extended only over the northern half of its unglaciated area (Gibbard and Cohen, 2015). This is also in agreement with the absence of Saalian and younger lacustrine sediments in the Belgian western plain (Bogemans et al., 2016).

The erosional processes that formed the others internal erosions interrupting the infills of the Fosses Dangeard are unknown. However, the flat morphology, the characteristics of the overlying seismic facies and the widespread spatial distribution of the youngest internal erosional surfaces (i.e. Ee2, Ed1, Ef1, Eb6 and Ea5) have some similarities with erosional surfaces associated with marine transgressions (e.g. Trincardi et al., 1994). It is thus possible that the formation of those erosional surfaces is related to a marine transgression that took place before the incision of the Lic.

The absence of sediment cores from Fosses A, B, C and D precludes establishing confidently the palaeo-environments in which the various infilling and scouring episodes composing the infill of the Fosses Dangeard occurred. This also prevents assessing their absolute ages. Currently, we can only state that the scouring-and-infilling episodes identified in the infill of the Fosses Dangeard postdates the waterfall phase that led to the breach of the Dover Strait dam, and that it predates the incision of the Lobourg inner channel (Lic).

6.6.2 Formation of the Lobourg Channel

The geomorphological analysis of the palaeovalleys located in the Dover Strait has revealed that the present-day Lobourg Channel is actually the combination of at least three phases of major valley incisions (Figure 6.10). The geomorphological features remaining from the first two do not allow the characterization of their original morphology and interrelationship, as the last one (Lic) truncates them both.

The three phases of valley incision are associated with sharp rectilinear slopes carved into chalk, as well as with longitudinal scours incised across the Wealden Beds and the infill of the Fosses Dangeard (Figure 6.4 and Figure 6.14). The Lic and LC2 are also associated with “tear-drop” streamlined islands carved into chalk. These features are indicative of highly erosive fluvial flows or high-magnitude flood flows, as their formation needs the flow to bifurcate, breaching through bedrock divide (see discussion in Collier et al., 2015). Features indicating extreme erosion are especially associated with the Lic. For instance, and contrarily to the previous two valley incisions, the Lic locally truncates the ridge formed at the seafloor by the Hythe Beds. In addition, the Lic runs across a range of bedrock sedimentary formations keeping a general box-shape cross-sectional morphology independently of the bedrock through which it cuts. This geomorphology is indeed similar to those found along high-energy fluvial systems and/or megaflood-eroded valleys (see Kehew and Lord, 1986; Wohl, 1993; Rains et al., 1993; Kale et al., 1996; Baynes et al., 2015). Finally, it comprises a range of bedrock erosions typically found in megaflood-eroded terrains, such as amphitheater-head channels, streamlined islands, longitudinal scours, linear ridge-and-groove bed-forms, etc. (Baker and Nummedal, 1978; Rains et al., 1993; Wohl, 1993; Lamb et al., 2007; Baker, 2009; Shaw, 2010; Baynes et al., 2015).

Taken together, the low topographic gradient across the Strait (see Lericolais et al., 2003), the geomorphology of the Lic and the erosional features carved within it indicate that at least this palaeovalley was most likely carved by a megaflood flow. Whether the formation of the previous two valley incisions LC1 and LC2 was also linked to megafloods is unknown. Nevertheless, the presence of streamlined islands associated with LC2 suggests that at least that palaeovalley was carved by highly erosive flows as well. Another feature pointing to possible high-magnitude flood flows occurring prior to the formation of the Lic are the ridge-and-groove set “La” (see Wohl, 1993), which seem to be truncated by the Lic (Figure 6.12). The occurrence of high-magnitude flood and/or intense fluvial erosions in the span between the formation of the Fosses Dangeard and the incision of the Lic is also attested by several internal erosional surfaces identified in Fosse B (i.e. Eb1 and Eb4). However, the relationship, if any, between the erosional event(s) that carved LC1 and LC2 and the internal erosions interrupting the infill of these palaeo-depression remains unknown.

The erosional features incised into the Lobourg Channel are not the same all along its length; they change depending on the stratigraphic units into which they have been carved. That is, ridge-and-groove bed-forms and “tear-drop” streamlined islands formed almost exclusively in the resistant chalk units (Figure 6.10 and Figure 6.12). Amphitheater-head palaeo-channels (Ch1 and Ch2) and linear scours (sc3) are, however, found in the geologically heterogeneous central Dover Strait (Figure 6.14). Finally, elongated and sinuous scours and elongated streamlined islands are distinctive of the southern Dover Strait. There, the Lobourg Channel cuts through the mudstones and clays composing the Wealden Beds, and the possibly poorly consolidated sediments infilling the Fosses Dangeard. These differences attest to the significant influence of the composition of the substratum on the morphology/type of bed-forms features that form during extreme erosional events.

Another important result from our geomorphological analysis is the identification of a SE–NW-oriented palaeovalley system (Pvs1) that appears to be tributary to the Lic (Figure 6.10). The fact that this system remains undisturbed across LC1 and LC2 implies that no further major river or high-magnitude flood flowed along the Lobourg Channel outside the Lic during or following the incision of Pvs1. That means that the Lic represents the last major fluvial erosion that imprinted the seafloor of the Dover Strait. This gives no indication on the actual age of its formation. However, it suggests that once the Lic was carved, all major rivers entering the Dover Strait from the southern North Sea basin were channeled through it.

6.6.3 Relative sequence of events in landscape evolution of the Dover Strait

This study has revealed evidence of several major scouring-and-infilling episodes in the Dover Strait, which took place prior to the formation of the Holocene tidal sand ridges. The main ones are, chronologically:

I. excavation of the Fosses Dangeard by plunge-pool erosion at the base of waterfalls;

II. breach of the Weald-Artois ridge, ending the waterfall phase and possibly generating a megaflood in the English Channel. Incision of some of the Channel palaeovalleys;

III. infilling of the Fosses Dangeard, which included 6 major scouring-and-infilling episodes in the center of the Strait. At least two of the scouring episodes were formed by intense fluvial scouring or flooding;

IV. first two phases of valley incision (LC1 and LC2) along the Lobourg Channel possibly by highly erosional fluvial system(s) and/or high-magnitude flood flows. It is however unclear whether the incisions of these features were contemporary to some of the internal erosional surfaces identified in Fosses A and B. That is, parts of step (3) and step (4) might have happened during the same subaerial exposure of the seafloor;

V. formation of the Lobourg inner channel (Lic) by a megaflood.

The absolute age of the various valley incisions identified in the Dover Strait are poorly constrained. The present study demonstrates though that the various erosional events that formed the Lobourg Channel postdate the formation of the Fosses Dangeard. However, it provides no information on the ages of the different erosional phases that formed them.

Even though sedimentary data and absolute dating are still lacking from this area, hypotheses can be made on the possible time of events by comparing with palaeogeographic reconstructions (see Figure 6.3). For instance, the waterfall phase that incised the Fosses Dangeard ended with the opening of the Dover Strait, which, according to palaeogeographic reconstructions, happened ~450 ka ago (e.g. Gibbard and Cohen, 2015). That age thus marks the last phase of waterfall incision in the Dover Strait, implying that the infilling-and-scouring episodes observed in the infill of the Fosses Dangeard and the various valley incisions that led to the formation of the Lobourg Channel would have had to occur during subsequent highstands and lowstands. The first major internal erosional surface observed in Fosses A and B (Ea1 and Eb1) was incised after a phase of infilling. It is thus possible that that and subsequent apparent fluvial erosional episodes observed in the infill of the Fosses Dangeard occurred during the next marine lowstand (i.e. the Saalian glaciation). Indeed, a major palaeo-river formed by the confluence of the southwestward-diverted Rhine–Meuse and Thames palaeo-rivers may have flowed through the Dover Strait at that time (Gibbard, 1995; Hijma et al., 2012). In addition, proglacial lakes, although smaller than the one that induced the formation of the Fosses Dangeard, might have formed again in the southern North Sea and other locations within the catchment of the Channel River during the Saalian glacial maximum (Meinsen et al., 2011; Gibbard and Cohen, 2015). The intense erosional episodes represented by erosional surfaces Eb1 and Eb4 might thus be related to lake-outburst floods generated by Saalian proglacial lakes and/or by episodes of rapid ice melting.

Concerning the formation of the Lic, the possible marine transgressional origin of the uppermost erosional surfaces identified in the infill of the Fosses Dangeard suggest that the

Lic, which truncates the sediments lying above that surface, may have been formed after a marine transgression. Hence, if Eb1 and Eb4 were formed during the Saalian glaciation, that marine transgression should represent the Eemian interglacial. The Lic might thus have been carved by megaflood(s) produced during the Last Glacial Maximum. This is consistent with the geomorphology of the Dover Strait, which suggest that the formation of the Lic and its tributaries (Pvs1) represent the last major subaerial erosional episode imprinted in the seafloor. If that hypothesis is correct, the origin of the megaflood(s) that carved the Lic might be related either to sudden increase(s) of water volumes of palaeo-rivers traversing the southern North Sea during episodes of rapid deglaciation (e.g. Toucanne et al., 2010; Toucanne et al., 2015), and/or lake-outburst flood(s) of the ice-marginal lake that, according to Sejrup et al. (2016), occupied part of the German–Danish Continental Shelf of the southern North Sea at 30–19 ka BP. We will elaborate and support this hypothesis with further evidence in next chapter.

Event	Associated erosional features	Possible age
Waterfall and plunge-pool erosion	Fosses Dangeard	Elsterian glacial max.
Breach of the ridge and first megaflood	Opening of the Dover Strait and possible incision of some of the Channel palaeovalleys	Elsterian glacial max.
Infilling-and-scouring episodes in the center of the Strait	Internal erosional surfaces in Fosses A and B.	Elsterian – Eemian
LC1 and LC2 valley incisions.	La?, E1, E2, sc1, sc2 and si0	Saalian glacial max.
Formation of the Lic by a megaflood and subsequent river erosion	Lb, Lc, si1, si1b, si2, si3, Ch1, Ch2 and sc3	Last Glacial Max.

Table 6.3. Possible correlations between erosional events and geomorphologic features identified in the Dover Strait. The possible timing of their occurrence is also indicated.

6.7 Conclusions

The detailed analysis of the high-resolution geophysical data available for this study provides new information on the formation and evolution of the Fosses Dangeard and the Lobourg Channel. In particular, we provide further geomorphological evidence supporting that the Fosses Dangeard are palaeo-plunge pools incised at the base of waterfalls, and the independence of that erosional event from (and older age than) the formation of the Lobourg Channel. Importantly, this study does not only corroborate that waterfalls (and so a ridge damming a lake in the southern North Sea) existed in the Dover Strait, but it also suggests that the wall overflowed by these retreated northward. That is, Fosses D, C, F and G indicate the position of the waterfall's wall when the cataract phase started, and the northern edges of Fosses A and E mark its position just before the end of that erosional phase. The morphology and spatial distribution of these palaeo-depressions is also consistent with a sudden breach of the Weald-Artois ridge and the generation of a megaflood flow in the English Channel.

Another important observation is the identification of only one phase of waterfall plunge-pool erosion in the Dover Strait; i.e. the one that incised the basal erosional surfaces of the Fosses Dangeard. This has important implications for palaeogeographic reconstructions of subsequent highstands and lowstands, as it indicates that the ridge damming the proglacial lake in the Dover Strait was opened, at least partially, when the waterfall phase ended. This validates recent palaeogeographic reconstructions, which suggested that lake-marginal lakes formed in the southern North Sea during subsequent marine lowstands did not extend this far southwest.

Our study also shows that the various depressions composing the Fosses Dangeard were subjected to a series of scouring-and-infilling episodes after their formation and before the occurrence of the youngest valley incision identified along the Lobourg Channel (i.e. the one that carved the Lic). More importantly, two of these indicate possible high-energy fluvial erosion and/or episodes of high-magnitude flooding.

The geomorphology of the Lobourg Channel has also revealed that this palaeovalley was shaped by, at least, three phases of intense valley incision. The various erosional features associated with the last one of these strongly suggest the occurrence of one or several megafloods in the Dover Strait. Our study thus corroborates the megaflood hypothesis postulated for the origin of the Lobourg Channel by previous studies, although only for its last phase of valley incision (i.e. formation of the Lic). Concerning the others (older) two valley incisions, we have not enough evidence to confidently associate them with megafloods. However, the geomorphological features associated with them indicate extreme erosion too.

The timing of the various subaerial erosional events that formed the Fosses Dangeard and Lobourg Channel are poorly constrained. Palaeogeographic reconstructions indicate that the opening of the Dover Strait may have happened ~450 ka ago, which implies that the Fosses Dangeard most likely formed during the Elsterian glacial maximum. The timing of the two intense, possibly fluvial and/or flooding, erosional episodes identified in the infills of the Fosses Dangeard (i.e. Eb1 and Eb4) and that of the two first valley incisions that shaped the Lobourg Channel (i.e. LC1 and LC2) are however unknown. With the data at hand, we can only state that they occurred in the span between the formation of the Fosses Dangeard (~450 ka BP) and the incision of the Lic. The absolute age of the latter erosional event is also unknown. However, our investigation has revealed that it may have taken place following a marine transgression. This added to the geomorphology of the seafloor in the Dover Strait, which indicates that the Lic represents the last major valley incision along the Strait, suggest that this inner channel was most likely carved during the Last Glacial Maximum.

In conclusion, the opening of the Dover Strait and the present-day geomorphology of its seafloor are the results of a sequence of extreme subaerial erosional events (e.g. megafloods), intense fluvial erosion and marine erosion. Importantly, the marine erosion is almost indistinguishable in the seafloor, where the erosional features carved by intense fluvial and/or flooding events stand out.

References

Antoine, P., Coutard, J-P., Gibbard, P., Hallegouet, B., Lautridou, J-P., and Ozouf, J-C. (2003), The Pleistocene rivers of the English Channel region, Journal of Quaternary Science, v.18, p. 227–243.

Alexandrowicz, Z., 1994, Geologically controlled waterfall types in the Outer Carpathians, Geomorphology, v. 9, p. 155–165.

Ashton, N., and Lewis, S., 2002, Deserted Britain: declining populations in the British Late Middle Pleistocene, Antiquity, v. 76, p. 388–396.

Ashton, N., and Hosfield, R., 2010, Mapping the human record in the British early Palaeolithic:

evidence from the Solent River system, Journal of Quaternary Sciences, v. 25, p. 737–753.

Ashton, N., Lewis, S. G., and Hosfield, R., 2011, Mapping the human record: population change in Britain during the Early Palaeolithic, p. 39-51. In: The ancient human occupation of Britain, (eds. Ashton, N., Lewis, J. E., and Stringer, C.).

Baker, V. R., and Nummedal, D., 1978, The Channeled Scabland; A Guide to the Geomorphology of the Columbia Basin, Washington, National Aeronautics and Space Administration.

Baker, V.R., 2009, The Channeled Scabland: A Retrospective, Annual Review of Earth and Planetary Sciences, v. 37, p. 393–411.

Balson, P.S., and D'Olier, B.D., 1989, Thames estuary solid geology, Sheet 51°N-00°, British Geological Survey, scale 1:250 000, 1 sheet.

Baynes, E. R. C., Attal, M., Dugmore, A. J., Kirstein, L. A. and Whaler, K. A., 2015, Catastrophic impact of extreme flood events on the morphology and evolution of the lower Jokulsa a Fjollum (northeast Iceland) during the Holocene, Geomorphology, v. 250, p. 422–436.

Bogemans, F., Baetaman, C., 2016, Incised Pleistocene valleys in the Western Belgium coastal plain: Age, origins and implications for the evolution of the Southern North Sea Basin. Palaeogeography, Palaeoclimatology, Palaeoecology, p. 456, v. 46–59.

Busschers, F. S., Weerts, H. J. T., Wallinga, J., Cleveringa, P., Kasse, C., De Wolf, H., and Cohen, K. M., 2005. Sedimentary architecture and optical dating of Middle and Late Pleistocene Rhine-Meuse deposits-fluvial response to climate change, sea-level fluctuation and glaciation, Netherlands Journal of Geosciences, v. 84, p. 25–41.

Busschers, F. S., Kasse, C., Van Balen, R. T., Vandenberghe, J., Cohen, K. M., Weerts, H. J. T., Wallinga, J., Johns, C., Cleveringa, P., and Bunnik, F. P. M., 2007. Late Pleistocene evolution of the Rhine-Meuse system in the southern North Sea basin: imprints of climate change, sea-level oscillation and glacio-isostacy, Quaternary Science Reviews, v. 26, p. 3216–3248.

Carr, S. J., Holmes, R. V. D., Van der Meer, J. J. M., & Rose, J., 2006, The Last Glacial Maximum in the North Sea Basin: micromorphological evidence of extensive glaciation, Journal of Quaternary Science, v. 21, p. 131–153.

Clark, P.U., Alley, R.B., Pollard, D., 1999. Northern hemisphere ice-sheet influences on global climate change, Science, v. 286, p. 1104–1111.

Clark, C.D., Gibbard, P.L. and Rose, J., 2004, Pleistocene glacial limits in England, Scotland and Wales, in Ehlers, J., and Gibbard, P.L., Eds., Developments in Quaternary Sciences. Elsevier, p. 47–82.

Cohen, K.M., Gibbard, P.L., and Weerts, H.J.T., 2014, North Sea palaeogeographical reconstructions for the last 1 Ma, Netherlands Journal of Geosciences – Geologie en Mijnbouw, v. 93, p. 1–23.

Collier, J. S., Oggioni, F., Gupta, S., García-Moreno, D., Trentesaux, A., De Batist, M., 2015, Streamlined islands and the English Channel megaflood hypothesis, Global and Planetary Change, v. 135, p. 190–206.

Crosby, A., Hamblin, R.J.O., Smart, J.G.O., and Bouysse, P., 1988, Dungeness-Boulogne solid geology, Sheet 50°N-00°, British Geological Survey, scale 1:250 000, 1 sheet.

De Clercq, M., Chademenos, V., Van Lancker, V., & Missiaen, T., 2016, A high-resolution DEM for the Top-Palaeogene surface of the Belgian Continental Shelf, Journal of Maps, v. 12, p. 1047–1054.

Destombes, J. P., Shephardthorn, E. R., Redding, J. H., and Morzadec-kerfourn, M. T., 1975. Buried valley system in the Strait of Dover, Philosophical Transactions of the Royal Society A-Mathematical Physical and Engineering Sciences, v. 279, p. 243–256.

Dyer, K.R., and Huntley, D.A., 1999. The origin, classification and modelling of sand banks and ridges. Continental Shelf Research, v. 19, p. 1285–1330.

Eilertsen, R. S., & Hansen, L., 2008, Morphology of river bed scours on a delta plain revealed by interferometric sonar, Geomorphology, v. 94, p. 58–68.

Ehlers, J., and Gibbard, P.L. (Eds.), 2004, Quaternary Glaciations—Extent and Chronology, Part I: Europe. Developments in Quaternary Science, vol. 2a. Elsevier, Amsterdam.

Gibbard, P.L., 1988, The history of the great northwest European nvers during the past three million years. Philosophical Transactions of the Royal Socity B, v. 318, p. 559–602

Gibbard, P.L., 1995, The formation of the Strait of Dover, Geological Society, London, Special Publications, v. 96, 15–26.

Gibbard, P.L., and Cohen, K.M., 2015, Quaternary evolution of the North Sea and the English Channel, Proceedings of the Open University Geological Society, v. 1, p. 63–74

Gibbard, P.L., and Lewin, J., 2016, Filling the North Sea Basin: Cenozoic sediment sources and river styles, Geologica Belgica, v. 19, p. 3–4.

Graham, A.G.C., Lonergan, L., Stoker, M.S., 2007, Evidence for Late Pleistocene ice stream activity in the Witch Ground basin, central North Sea, from 3D seismic reflection data, Quaternary Science Reviews, v. 26, p. 627–643.

Gupta, S., Collier, J. S., Palmer-Felgate, A., and Potter, G., 2007, Catastrophic flooding origin of shelf valley systems in the English Channel. Nature, v.448, p. 342-345

Hamblin, R.J.O., Crosby, A., Balson, P.S., Jones, S.M., Chadwick, R.A., Penn, I.E., and Arthur, M.J., 1992, United Kingdom offshore regional report: the geology of the English Channel. London: HMSO for the British Geological Survey, 106 p.

Hancock, G. S., and Anderson, R. S., 2002, Numerical modeling of fluvial strath-terrace formation in response to oscillating climate. Geological Society of America Bulletin, v. 114, p. 1131–1142.

Hijma, M. P., Cohen, K. M., Roebroeks, W., Westerhoff, W. E., and Busschers, F. S., 2012, Pleistocene Rhine–Thames landscapes: geological background for hominin occupation of the southern North Sea region. Journal of Quaternary Science, v. 27, p. 17–39.

Hopson, P.M., Wilkinson, I.P. and Woods, M.A., 2008, A stratigraphical framework for the Lower

Cretaceous of England, Research Report RR/08/03. British Geological Survey, Keyworth.

James, J.W.C., Guennoc, P., Harrison, M., Le Bot, S., Philpott, S., Vinchon, C., Bee, E., Simien, F., Janjou, D., Garlan, T., Trentesaux, A., Mahieux, G., Briet D. and, Augris C., 2002, GEOSYNTH: A synthesis of the geology and sediments of the Dover Strait and its hinterland. CR/02/078, CDROM, British Geological Survey.

Kale, V.S., Baker, V.R., and Mishra, S., 1996, Multi-channel patterns of bedrock rivers: An example from the central Narmada basin, India. Catena, v. 26, p. 85–98.

Kehew, A.E. and Lord, M.L., 1986, Origin and large-scale erosional features of Glacial-Lake spillways in the northern Great-Plains. Geological Society of America Bulletin, v. 97, p. 162–177.

Kristensen, T. B., Huuse, M., Piotrowski, J. A., and Clausen, O. R., 2007, A morphometric analysis of tunnel valleys in the eastern North Sea based on 3D seismic data, Journal of Quaternary Science, v. 22, p. 801–815.

Kjemperud, A. T., and Fjeldskaar, W., 1992, Pleistocene glacial isostasy—implications for petroleum geology. Tectonic modelling and its application to petroleum geology. Norwegian Petroleum Society, Special Publication, v. 1, p. 187-195

Lamb, M.P., Howard, A.D., Dietrich, W.E. and Perron, J.T., 2007, Formation of amphitheater-headed valleys by waterfall erosion after large-scale slumping on Hawaii. Geological Society of America Bulletin, v. 119, p.805–822.

Lamb, M.P., 2008, Formation of Amphitheater-Headed Canyons, PhD dissertation, University of California, Berkeley, 311 p.

Lee, J. R., Busschers, F. S., and Sejrup, H. P., 2012, Pre-Weichselian Quaternary glaciations of the British Isles, The Netherlands, Norway and adjacent marine areas south of 68 N: implications for long-term ice sheet development in northern Europe. Quaternary Science Review, v. 44, p. 213–228.

Lericolais, G., Auffret, J.-P., and Bourillet, J.-F., 2003, The Quaternary Channel River: seismic stratigraphy of its palaeovalleys and deeps, Journal of Quaternary Science, v. 18, p. 245–260.

Lonergan, L., Maidment, S. C., & Collier, J. S., 2006, Pleistocene subglacial tunnel valleys in the central North Sea basin: 3-D morphology and evolution, Journal of Quaternary Science, v. 21, p. 891–903.

Mansy, J.-L., Manby, G.M., Averbuch, O., Everaerts, M., Bergerat, F., Van Vliet-Lanoe, B., Lamarche, J. and Vandycke, S., 2003, Dynamics and inversion of the Mesozoic Basin of the Weald-Boulonnais area: role of basement reactivation, Tectonophysics, v. 373, p. 161–179.

Meinsen, J., Winsemann, J., Weitkamp, A., Landmeyer, N., Lenz, A., and Dölling, M., 2011, Middle Pleistocene (Saalian) lake outburst floods in the Münsterland Embayment (NW Germany): impacts and magnitudes, Quaternary Science Reviews, v. 30, p. 2597–2625.

Meijer, T. & Preece, R. C., 1995, Malacological evidence relating to the insularity of the British Isles during the Quaternary, Geological Society, London, Special Publications, v. 96, p. 89-110.

Mellett, C., Hodgson, M., Plater, A., Mauz, B., Selby, I. & Lang, A., 2013, Denudation of the continental shelf between Britain and France at the glacial–interglacial timescale, Geomorphology, v. 203, p. 79–96.

Mortimore, R.N., 2011. A chalk revolution: what have we done to the Chalk of England?, Proceedings of the Geologists' Association, v. 122, p. 232–297.

Pagliara, S., Hager, W. H., and Unger, J., 2008, Temporal evolution of plunge pool scour, Journal of Hydraulic Engineering, v. 134, p. 1630–1638.

Patton, H., Hubbard, A., Andreassen, K., Auriac, A., Whitehouse, P. L., Stroeven, A. P., Shackleton.C., Winsborrow, M., Heyman, J., and Hall, A. M., 2017, Deglaciation of the Eurasian ice sheet complex, Quaternary Science Reviews, v. 169, p. 148–172.

Praeg, D., 2003, Seismic imaging of mid-Pleistocene tunnel-valleys in the North Sea Basin—high resolution from low frequencies, Journal of Applied Geophysics, v. 53, p. 273– 298.

Preece, R. C., 1995, Island Britain: a Quaternary Perspective, v. 96, Geological Society Special Publication.

Radley, J.D., and Allen, P., 2012, The Wealden (non-marine Lower Cretaceous) of the Weald Sub-basin, southern England, Proceedings of the Geologists' Association, v. 123, p. 245–318.

Rains, B., Shaw, J., Skoye, R., Sjogren, D., and Kvill, D., 1993, Late Wisconsin subglacial megaflood paths in Alberta. Geology, v. 21, p. 323–326.

Reynaud, J-Y., Tessier, B., Auffret, J-P., Berné, S., De Batist, M., Marsset, T., and Walker, P., 2003, The offshore Quaternary sediment bodies of the English Channel and its Western Approaches, Journal of Quaternary Science, v. 18, p. 361–371.

Roep, T. B., Holst, H., Vissers, R. L. M., Pagnier, H., and Postma, D., 1975, Deposits of southward-flowing, pleistocene rivers in the channel region, near Wissant, NW France. Palaeogeography, Palaeoclimatology, Palaeoecology, v. 17, p. 289–308.

Sejrup, H. P., Clark, C. D., and Hjelstuen, B. O., 2016, Rapid ice sheet retreat triggered by ice stream debuttressing: Evidence from the North Sea. Geology, v. 44, p. 355–358.

Shaw, J., 2010, In defence of the meltwater (megaflood) hypothesis for the formation of subglacial bedform fields, Journal of Quaternary Science, v. 25, p. 249–260.

Smith, A.J., 1985, A catastrophic origin for the palaeovalley system of the eastern English Channel. Marine Geology, v. 64, p. 65–75.

Strasser, M., Schindler, C., and Anselmetti, F. S., 2008. Late Pleistocene earthquake-triggered moraine dam failure and outburst of Lake Zurich, Switzerland. Journal of Geophysical Research: Earth Surface, v. 113, doi: 10.1029/2007JF000802

Stuart, A. J., 1995, Insularity and Quaternary vertebrate faunas in Britain and Ireland, Geological Society, London, Special Publications, v. 96, p. 111-125.

Sturt, F., Garrow, D. and Bradley, S., 2013, New models of North West European Holocene palaeogeography and inundation, Journal of Archaeological Science, v. 40, p. 3963–3976.

Sutcliffe, A. J., 1995, Insularity of the British Isles 250 000–30 000 years ago: the mammalian, including human, evidence, Geological Society, London, Special Publications, v. 96, p. 127-140.

Toucanne, S., Zaragosi, S., Bourillet, J. F., Gibbard, P. L., Eynaud, F., Giraudeau, J., Turon, J.L., Cremer, M., Cortijo, E., Martinez, P., and Rossignol, L., 2009a, A 1.2 Ma record of glaciation and fluvial discharge from the West European Atlantic margin. Quaternary Science Reviews, v. 28, p. 2974–2981.

Toucanne, S., Zaragosi, S., Bourillet, J.F., Cremer, M., Eynaud, F., Van Vliet-Lanoë, B., Penaud, A., Fontanier, C., Turon, J.L., Cortijo, E. and Gibbard, P.L., 2009b, Timing of massive 'Fleuve Manche' discharges over the last 350 kyr: insights into the European ice-sheet oscillations and the European drainage network from MIS 10 to 2, Quaternary Science Reviews, v. 28, p. 1238–1256.

Toucanne, S., Zaragosi, S., Bourillet, J. F., Marieu, V., Cremer, M., Kageyama, M., Van Vliet-Lanoë, B., Eynaud, F., Turon, J-L, and Gibbard, P. L., 2010, The first estimation of Fleuve Manche palaeoriver discharge during the last deglaciation: evidence for Fennoscandian ice sheet meltwater flow in the English Channel ca 20–18ka ago, Earth and Planetary Science Letters, v. 290, p. 459–473.

Toucanne, S., Soulet, G., Freslon, N., Jacinto, R. S., Dennielou, B., Zaragosi, S., Eynaud, F., Bourillet, J-F., and Bayon, G., 2015, Millennial-scale fluctuations of the European Ice Sheet at the end of the last glacial, and their potential impact on global climate. Quaternary Science Reviews, v. 123, p. 113-133.

Trincardi, F., Correggiari, A., and Roveri, M., 1994, Late Quaternary transgressive erosion and deposition in a modern epicontinental shelf: the Adriatic semienclosed basin. Geo-Marine Letters, v. 14, p. 41-51.

Turner, C., 2000. The Eemian interglacial in the North European plain and adjacent areas. Nethederland Journal Geoscience, v. 79, p. 217–231.

Whipple, K.X., Hancock, G.S., and Anderson, R.S., 2000. River incision into bedrock: Mechanics and relative efficacy of plucking, abrasion, and cavitation, Geological Society of America Bulletin, v. 112, p. 490–503.

Wohl, E.E., 1993, Bedrock Channel Incision along Piccaninny Creek, Australia, Journal of Geology, v. 101, p. 749–761.

Zagwijn, W.H., 1989. The Netherlands during the Tertiary and the Quaternary: a case history of Coastal Lowland evolution. In: Coastal Lowlands, Springer Netherlands, p. 107–120.

Chapter 7

Geomorphology of the southern North Sea palaeovalley systems. Middle–Late Pleistocene formation and evolution.

Chapter 7 – Geomorphology of the southern North Sea palaeovalley systems. Middle–Late Pleistocene formation and evolution.

Abstract

Palaeogeographic reconstructions suggest that ice sheets extended over large parts of northern Europe during major Pleistocene glacial maxima, isolating the southern North Sea from the North Atlantic Ocean over long periods. These particular palaeogeographic/palaeo-climatic settings induced important modifications in the drainage systems of northwestern Europe. In particular, northwestern European drainage systems extended across the then emerged southern North Sea, forming large river systems and lakes. Even though the palaeo-hydrographic network on land is relatively well understood, the characteristics of the various drainage systems traversing the southern North Sea during the various glacial stages remain poorly constrained due to the lack of detailed offshore geomorphological studies. The present study aims to better constrain the distribution and evolution of the various palaeo-drainage systems that traversed the southern North Sea during major Middle–Late Pleistocene glacial periods. With that purpose, we have characterized the morphology and interrelationship of the main exposed and buried palaeovalleys and palaeo-depressions imprinted on the southern North Sea by combining extensive bathymetric and seismic reflection datasets.

The various palaeovalleys and palaeo-depressions imprinted on the southern North Sea demonstrates the occurrence of a series of fluvial and fluvio-glacial erosional episodes. Despite the absence of absolute dating, we have been able to tentatively associate the formation of most of these features with specific Pleistocene glaciations thanks to their geomorphology, extents and relative ages. This has permitted to test a series of hypotheses previously postulated, but never properly demonstrated, as well as to shed some light on the origin of several enigmatic erosional features incised in this area. In particular, the geomorphological analysis of the Outer Thames Estuary has revealed a series of, possibly Elsterian, palaeo-depressions similar to those observed in glacially eroded palaeo-terrains. This indicates that the Elsterian ice sheet may have extended further southward than previously thought. We have also found evidence corroborating the Elsterian southward diversion of the Thames–Medway palaeo-river and the southward migration of the eastern end of its watercourse during subsequent Pleistocene glacial periods. Furthermore, the morphology and spatial distribution of palaeovalleys in the southern North Sea supports the model in which the Rhine–Meuse and Thames palaeo-river systems converged together into the Axial Channel, which entered the English Channel through the Lobourg Channel, at least, during the Weichselian glacial maximum. Our study also corroborates that important water input coming from northwestern Germany joined that system. These palaeo-rivers incised a prominent palaeovalley in the seafloor; in this study referred to as "North Axial Channel". Importantly, we have found evidence suggesting that the incision of the Lobourg Inner Channel (Lic) described in the previous chapter of this dissertation may be the southwestward continuation of similar inner channels identified within the North Axial Channel and the Axial Channel. Correlations with palaeogeographic reconstructions suggest that the megaflood event that possibly carved these inner channels, which represents the last phase of valley incision along the Lobourg–Axial palaeovalley, may have occurred during the last stages of the Last Glacial Maximum.

Keywords: Geomorphology; Palaeogeography; North Sea; Pleistocene

Author contributions: Marc De Batist (M.D.B.) and David García–Moreno (D.G-M.) conceived, designed and coordinated the study. Maikel De Clercq (M.D.C.) gathered and interpreted 1980–2015 Belgian Continental Shelf geophysical dataset with contributions from D.G-M. and Vasileios Chademenos (V.C.). V.C. gathered and processed borehole information from Belgian Continental Shelf. M.D.C., D.G-M., V.C., Tine Missiaen (T.M.) and Oscar Zurita Hurtado (O.Z.H.) designed, coordinated and conducted RV Simon Stevin geophysical surveys on the Belgian Continental Shelf and Outer Thames Estuary. O.Z.H. processed seismic reflection data. V.C. produced velocity model for time–depth conversion. D.G.M. gathered Eastern Anglian and Outer Thames Estuary geophysical data. D.G-M. compiled all data in one dataset. D.G.M. analyzed and interpreted the data with contributions from M.D.C. D.G-M. wrote the paper with contributions from Kris Vanneste (K.V.), M.D.B. and M.D.C.

7.1. Introduction

The Pleistocene evolution of major drainage systems discharging into the southern North Sea is relatively well known on land (Bridgland and D'Olier 1995; Lewis et al., 2004; Busschers et al. 2005; Busschers et al. 2007; Bateman et al., 2008; Bateman et al., 2011; Peeters et al. 2015; Peeters et al. 2016). Nonetheless, their continuations through the submarine southern North Sea during major marine lowstands are poorly constrained. Indeed, no specific geomorphological study on the various palaeovalleys traversing that area has been undertaken prior to this study. Previous reconstructions of the southern North Sea hydrographic network during Pleistocene lowstand intervals are mainly based on the combination of on-land studies with local offshore surveys and reconstructions of the ice-sheet expansions at each Pleistocene Glacial maximum (e.g. Henriet et al., 1989; Liu et al., 1992; Gibbard, 1995; Hijma et al. 2012; Cohen et al., 2014; Gibbard and Cohen, 2015).

Present-day paleo-geographic reconstructions postulate that the Scandinavian and British–Irish ice sheets merged across the central and northern North Sea during the last three major Pleistocene glacial maxima (e.g. Clark et al., 2004; Ehlers and Gibbard, 2004; Sejrup et al., 2009, Lee et al., 2012; Sejrup et al., 2016). This caused the isolation of the then emerged southern North Sea from the North Atlantic Ocean (e.g. Gibbard, 1995; Gibbard and Cohen, 2015). At the Elsterian glacial maximum (i.e. MIS 12; ~450 ka BP), this isolation appears to have induced the formation of a proglacial lake dammed in the north by the coalescent ice sheets and in the south by the present-day Dover Strait, which was still closed at that time (See Chapters 5 and 6; e.g. Gibbard, 1988; Gibbard and Cohen, 2015; Gibbard and Lewis, 2016). The main outflow of that proglacial lake seems to have entered the English Channel across the Dover Strait, firstly, as large waterfalls as the lake overtopped the chalk ridge damming the lake at that location and, lastly, as a high-magnitude flood flow that most probably evolved into a major river.

Apart from the presence of a proglacial lake dammed at the Dover Strait, little is known about the drainage pattern within the southern North Sea before and during the Elsterian glacial period. Only a W–E-oriented palaeovalley extending eastward from the present-day Stour–Orwell River's mouth and a couple of elongated deeps carved in the Outer Thames Estuary (i.e. The Inner Gabbard Deeps) have been tentatively associated with these periods (Figure 7.1

and Figure 7.2; e.g. Bridgland and D'Olier, 1995; Emu Ltd and University of Southampton, 2009).

The W–E-oriented palaeovalley is typically associated with the pre-Elsterian Thames–Medway palaeo–river system. According to Bridgland and D'Olier (1995), among others, the Thames–Medway palaeovalley system entered the southern North Sea through that watercourse until the Elsterian glacial maximum, when its eastern part was diverted southward. Those authors have also proposed that the whole river system migrated southward during that glacial period, establishing itself along its present-day on-land watercourse (e.g. Gibbard, 1995; Bridgland and D'Olier, 1995).

The origin of the Inner Gabbard Deeps (IGD) is uncertain. The authors of the report by Emu Ltd and University of Southampton (2009) hypothesized that these deeps might have been formed by glacial erosion. However, they did not provide any evidence in support of that hypothesis. More recently, Hijma et al. (2012) suggested that fluvial systems may have flowed through that area 150 ka and/or 70 ka ago, which may also have contributed to the incision of (or formed) these palaeo-depressions. Nonetheless, Hijma et al. (2012) did not provide any evidence proving that hypothesis either.

According to, among others, Gibbard and Cohen (2015), at the Saalian glacial maximum (i.e. MIS 6; 175–140 ka BP), the southern North Sea was again inundated by a proglacial lake, which evolved into a continuous major river system around 150–140 ka BP, comprising the Meuse, the Rhine, the Thames and the Channel palaeo-river system (Gibbard, 1995; Toucanne et al. 2009b; Gibbard and Cohen, 2015). This evolution of the drainage pattern has however never been properly demonstrated, as few remnants from those palaeo-drainage systems have been identified in the southern North Sea. Currently, only a SW-trending scarp sub-cropping underneath the Holocene cover across the Belgian Continental Shelf (BCS) has been tentatively associated with that period. This scarp is interpreted as part of the western palaeo-slope of the Saalian Rhine–Meuse palaeo-river system (Mathys, 2009; Hijma et al., 2012).

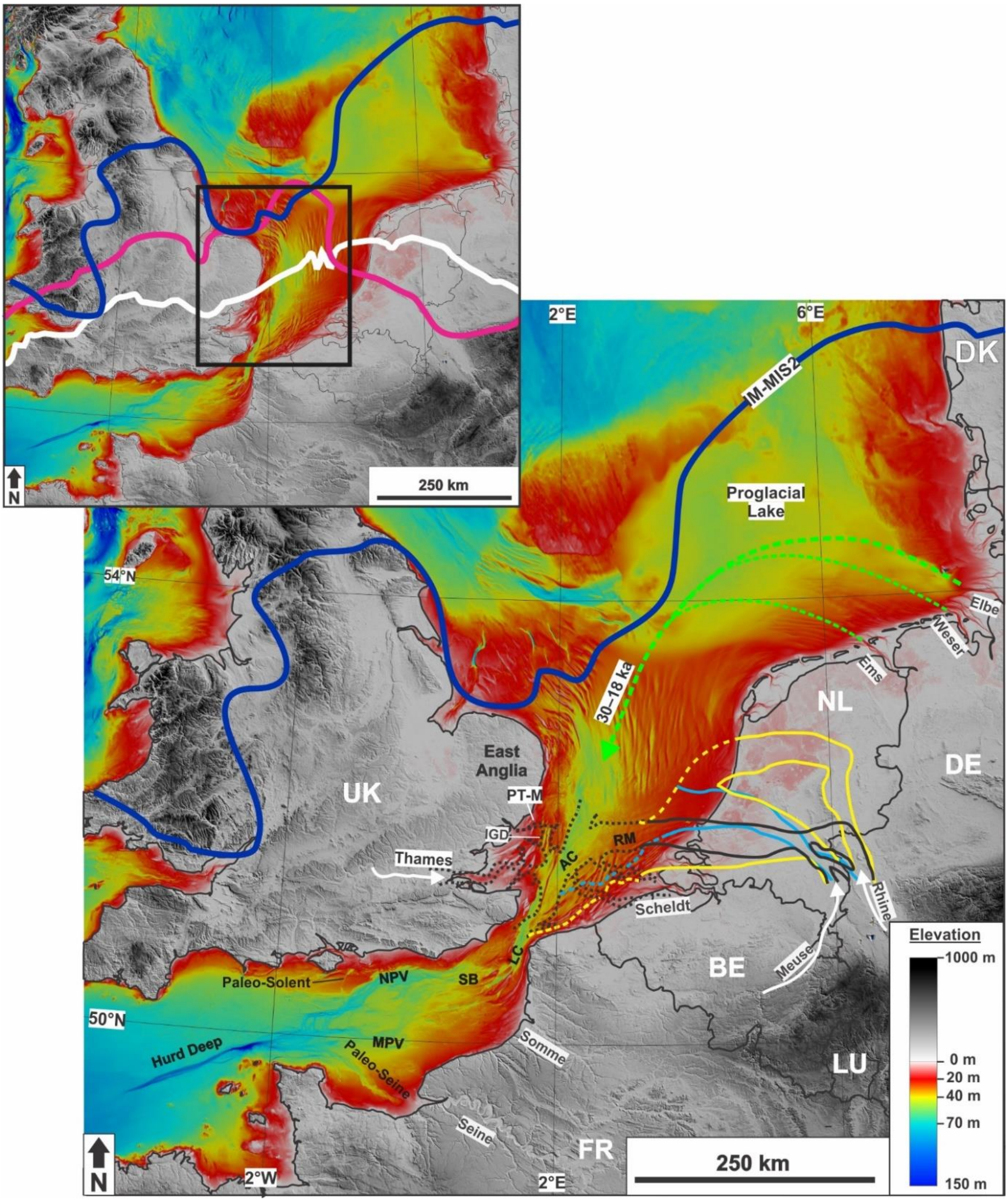

Figure 7.1. Main exposed and possible buried palaeovalleys imprinted on the English Channel and southern North Sea (after Hijma et al., 2012; Mellett et al., 2013). Dotted and dashed lines: hypotheses; continues lines: based on geological and/or geomorphological evidence. Light blue lines, Yellow and black lines represent the palaeo-watercourses of the Rhine–Meuse palaeo-river at 150 ka BP, at 80–35 ka BP and at 30–25 ka BP, respectively. After Busschers et al. (2007); Peeters et al. (2015) and Hijma et al. (2012). RM: possible continuation of the Rhine–Meuse palaeovalley at 30-25 ka (Hijma et al., 2012); IM: ice maximum extent; AC: Axial Channel; IGD: Inner Gabbard Deeps; PT–M: pre-Elsterian Thames-Medway palaeovalley. Light green dashed lines indicate the hypothetical southwestward deviation of northwestern German rivers (after Toucanne et al., 2010). Proglacial Lake: possible location of an ice-marginal lake that, according to Sejrup et al.

(2016), occupied that area 30–19 ka ago. Major palaeovalleys composing the Channel palaeovalley system are labelled as: LC: Lobourg Channel; SB: South Basserelle palaeovalley; NPV: Northern palaeovalleys; MPV: Median palaeovalley; and Hurd Deep. Inset map: North Sea southern maximum extents of ice sheets during Elsterian, Saalian and Weichselian glacial maxima (white, pink and dark blue lines, respectively), after Ehlers and Gibbard (2004), Lee et al. (2012) and Sejrup et al. (2016). Black rectangle: area shown in Figure 7.2 to Figure 7.4 and in Figure 7.15 to Figure 7.18. Bathymetry: EMODnet dataset (230 m cell size). Topographic information: SRTM worldwide elevation data (3-arc-second resolution). Projections: UTM, zone 31N. Horizontal datum: WGS84; SRTM vertical datum: EGM96; Bathymetric vertical datum: LAT.

During the Weichselian glaciation (also known as Last Glacial period; MIS 5c–2; 110–12 ka BP), the British–Irish and Scandinavian ice-sheets merged again across the North Sea at 30–19 ka BP (e.g. Carr et al., 2006; Sejrup et al., 2016). In that context, northwestern German rivers previously running northward, such as the Elbe, the Weser and the Ems Rivers, were deviated southwestward (Figure 7.1; e.g. Toucanne et al., 2010; Toucanne et al., 2015; Sejrup et al., 2016). These palaeo-rivers appear to have converged with the also southwestward diverted Rhine–Meuse and Thames palaeo-river systems into the Axial Channel (Toucanne et al. 2010; Hijma et al., 2012; Toucanne et al. 2015; Sejrup et al., 2016). The Axial Channel purportedly continued southwestward connecting with the Channel palaeo-river through the Lobourg Channel (Figure 7.1; e.g. Hijma et al., 2012; Mellett et al., 2013).

Even though these palaeogeographic reconstructions are widely accepted, no geomorphological evidence has been yet presented to validate them. In addition, the palaeo-drainage system derived from the southward diversion of the German rivers discharging into the southern North Sea and its relationship with the Rhine–Meuse–Thames–Axial–Channel palaeo-river system is unknown. Previous studies have theorized that either these rivers converged into a southwestward-trending major river (Figure 7.1; Toucanne et al., 2010; Toucanne et al., 2015) or they formed an ice-marginal lake (Sejrup et al. 2016; Patton et al., 2017). In any case, both hypotheses agree that the palaeo-river formed by the lake outflow, or by the merged palaeo-river system, ran toward the Dover Strait until 19–18 ka BP, converging with the Meuse, Thames and Rhine palaeo-rivers in the Axial Channel (Toucanne et al., 2010; Toucanne et al., 2015; Sejrup et al. 2016; Patton et al., 2017).

In this study, we combine several sets of seismic reflection and bathymetric data in order to characterize the geomorphology and interrelationship of the various Pleistocene palaeovalleys imprinted on the southern North Sea. With this, we aim to test the different palaeogeographic models proposed for the evolution of the major southern North Sea drainage systems during Middle–Late Pleistocene glaciations. In particular, we will focus on the local geomorphology of the Thames and Outer Thames Estuary area, and the geomorphology of the regional Rhine–Meuse–Thames–Axial–Lobourg palaeovalley system. We will also investigate the possible intersection of the latter by southward diverted German rivers during the Last Glacial period.

7.2. Geological setting

The North Sea is a continental shelf sea located in between northwestern European continental coast, the Scandinavian Peninsula and Great Britain. Our study focuses on the part of the southern North Sea that was free of ice during the last Pleistocene glacial period, part of which was not glaciated during the previous two glaciations either (Figure 7.1).

Because this study focuses on subaerial erosional features formed during and after the Elsterian glacial maximum, we refer to pre-Elsterian (i.e. deposited before MIS 12) geological formations as “substratum”. The fact that these erosional features were carved into Neogene or early Quaternary sediments is indeed irrelevant for the purpose of this study.

The pre-Quaternary substratum is characterized by NE-dipping, NNW–striking Neogene and Paleogene sedimentary formations (Figure 7.2; e.g. Balson et al., 1991). These include more or less consolidated clays, silts and sands, which are locally faulted and folded (e.g. Balson and D’Olier, 1989; Balson et al. 1991; Le Bot et al., 2003). Cretaceous and Jurassic formations are restricted to the Dover Strait and to the Spur area (Figure 7.2). In the Dover Strait, pre-Quaternary formations outcrop at the seafloor. These formations include upper Cretaceous chalk in the north and Lower Cretaceous sandstones, mudstones, sands and clays in the south (e.g. Hamblin et al., 1992). In the Spur area, Cretaceous formations sub-crop beneath the Quaternary cover together with Jurassic and Triassic clays, siltstones, sandstones, limestones and evaporites (e.g. Gaunt et al., 1985; Henriet et al., 1989; Tappin, 1991).

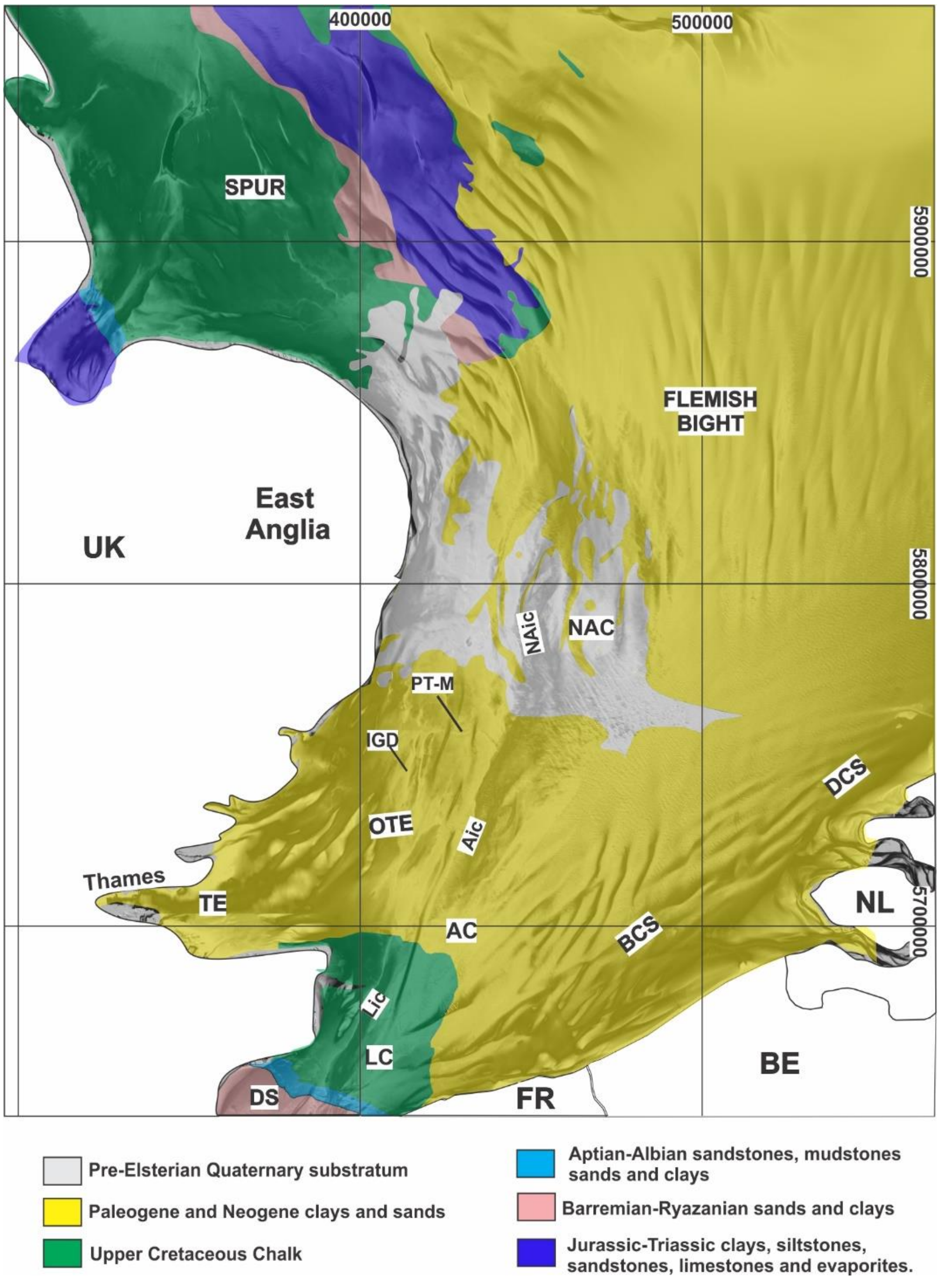

Figure 7.2. Geological map of the southern North Sea and northern Dover Strait showing the spatial distribution of pre-Elsterian sedimentary formations. After BGS maps produced by Cameron et al. (1984b), Gaunt et al. (1985), Zalasiewicz and Balson (1985), Cameron et al. (1987b), Blason and D'Olier (1989), Blason et al. (1991). TE: Thames Estuary; OTE: Outer Thames Estuary; PT-M: Pre-Elsterian Thames–Medway palaeovalley; DS: Dover Strait; NAC: North Axial Channel; Aic: Axial Inner Channel; NAic: North Axial Inner Channel; LC: Lobourg Channel; Lic: Lobourg Inner Channel; IGD: Inner Gabbard Deeps; BCS: Belgian Continental Shelf; DCS: Dutch Continental Shelf. For this and next figures projections: UTM zone 31N/WGS84/meters.

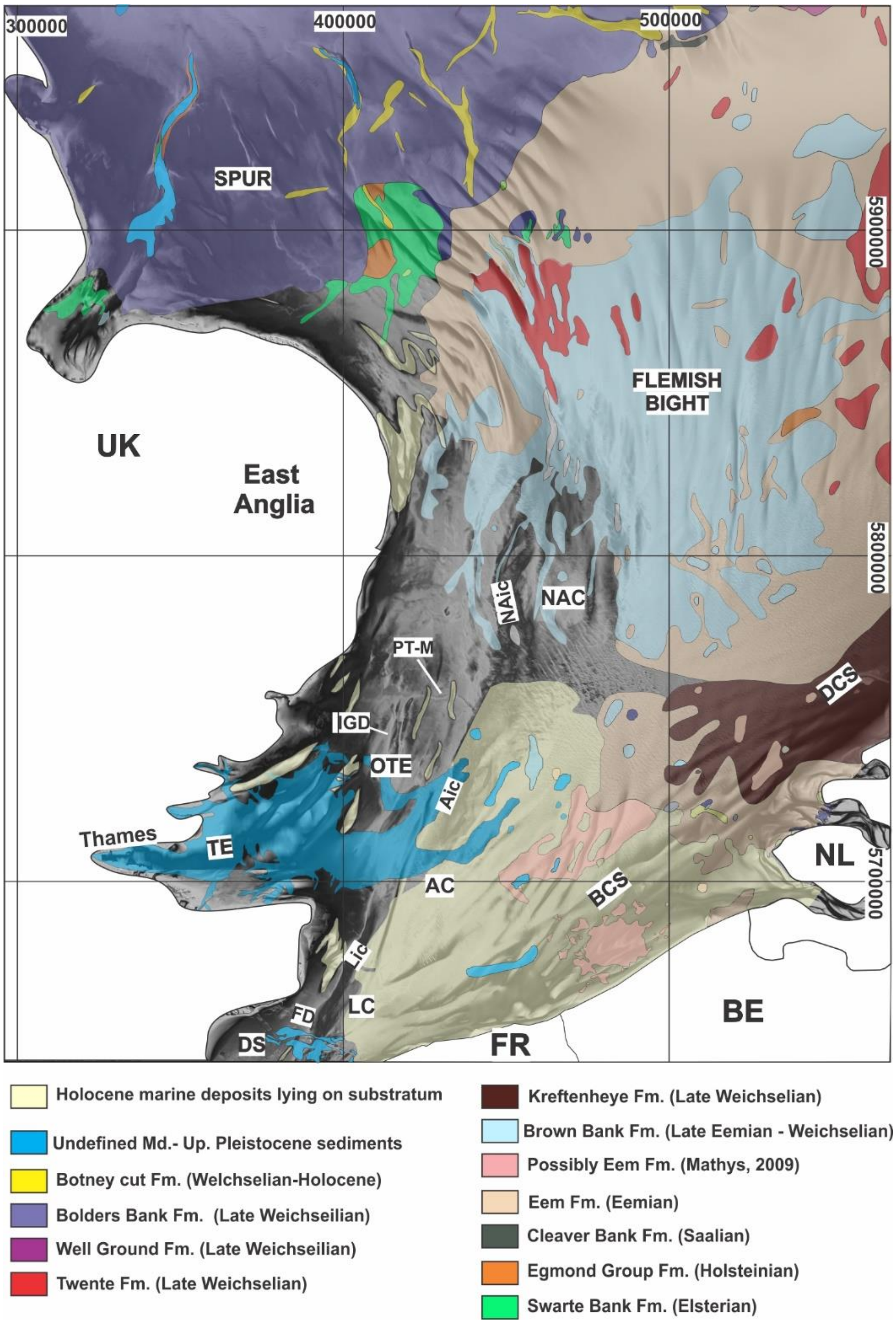

Figure 7.3. Post-Elsterian Quaternary sedimentary formations outcropping at the seafloor or sub-cropping below Holocene sediments. After BGS maps produced by Cameron et al. (1984a), Cameron et al. (1986), Blason and D'Olier (1990), Tappin (1991), and Laban et al. (1992). Major geomorphological features are indicated (labels already defined in Figure 7.2). Undefined Middle–Late Pleistocene sediments in TE: infill of Thames buried palaeovalleys. Note that channels Lic, Aic and NAic are mostly carved into substratum.

Pre-Elsterian Quaternary substratum includes marine, locally shelly, clays and sands, which evolve upwards to fluvial sands with clay intercalations (Cameron et al., 1984a). These units reach locally hundreds of meters in thickness (Cameron et al., 1984a). In the study area, these units outcrop at the seafloor along a belt a few tens of kilometers wide, extending parallel to East Anglia coastline, from the Outer Thames Estuary to the Spur area (Figure 7.2; e.g. Balson and D'Olier, 1989; Hamblin et al., 1992; James et al., 2002).

The Middle–Late Pleistocene sedimentary units reach their maximum thickness along the Dutch Continental Shelf, Flemish Bight and Spur areas (see Chapter 2). Elsewhere, these sediments are rather localized. For instance, in the Dover Strait, Pleistocene sediments are limited to the infills of the Fosses Dangeard (Figure 7.3; e.g. James et al., 2002). The absolute age of these sediments is however unknown. Their age is inferred to be younger than ~450 ka BP, when the Fosses Dangeard purportedly formed, and older than the last major valley incision identified along the Lobourg Channel, the age of which is currently poorly constrained (see Chapter 6).

The prominent palaeovalley that traverses the Dover Strait (i.e. The Lobourg Channel) is mostly unfilled. The Lobourg Channel forms part of the hundreds-of-kilometers long Channel palaeovalley system, also known as the Channel River (Figure 7.1; e.g. Mellett et al., 2013; Collier et al., 2015). This prominent channel appears to connect with the Rhine–Meuse and Thames palaeovalleys through the Axial Channel (e.g. Hijma et al., 2012). The Axial Channel, on the other hand, is partially infilled with Middle–Late Pleistocene sediments in the Outer Thames Estuary (see Figure 7.3), which may also be infilling the Thames buried palaeovalleys (see also Liu et al., 1993; Bridgland and D'Olier, 1995). Otherwise, the Axial Channel, similarly to its southward continuation (i.e. the Lobourg Channel), present little to no Middle–Late Pleistocene cover. The Rhine–Meuse palaeovalley is also infilled near the Netherlands coast by Late Weichselian and Eemian sediments. However, its possible westward continuation through the Belgian Continental Shelf (Figure 7.1) appears to be only covered by Holocene sediments (Figure 7.2).

The Quaternary geology of the rest of the study area comprises a range of Middle–Late Pleistocene sedimentary units locally covered by Holocene marine sands forming more or less extensive linear sandbank and dune fields (Figure 7.4; e.g. Dyer and Huntley, 1999). The main units composing the post-Elsterian Quaternary sediments deposited in this area are described in Chapter 2. The extents of their outcrop in the southern North Sea are plotted in Figure 2.3 and Figure 7.3.

Holocene marine sediments cover the entire study area. In large parts of the Axial Channel, East Anglia, the Spur area and the Dover Strait, this cover is rather thin, ranging between 0–2 m (e.g. James et al., 2002; Tappin, 1991; Cameron et al., 1984b; Balson and D'Olier, 1990). Holocene sedimentary units reach their maximum thicknesses along sandbank and dune fields (e.g. Houbolt, 1968; Caston and Stride, 1973; Dyer and Huntley, 1999). These include tens-of-kilometers long, several kilometers wide, sub-parallel elongated sandbanks, and major (kilometer-scale) and minor (tens to hundreds of meters long) sand dunes (Figure 7.4; Dyer and Huntley, 1999). The formation of these major sedimentary features is associated with the Holocene marine transgression and subsequent submarine currents (see Dyer and Huntley, 1999).

7.3. Methodology

In the present study, we combine several sets of 2D seismic-reflection data with various bathymetric datasets of different resolutions. Seismic reflection data include two main datasets; i.e. ~700 line-kilometers of 2D high-resolution seismic-reflection profiles acquired from the Outer Thames Estuary (OTE) in 2008, and ~7000 line-kilometers of 2D seismic-reflection data with various resolutions, comprising data acquired from the Belgian Continental shelf since the 1980's.

The bathymetric data comprises six datasets; these are: the BCS dataset, gridded at 20 m bin size; the Dutch Continental Shelf (DCS) dataset, gridded at 25 m bin size; the Dover Strait dataset, gridded at 40 m bin size; the Gabbard Sandbank dataset, gridded at 10 m bin size; the OTE dataset, gridded at 1 m bin size; the East Coast REC dataset, gridded at 1 m bin size; and the Emodnet Bathymetric DTM, gridded at 230 m bin size (see Figure 3.6). These datasets are described in detail in Chapter 3 of this dissertation. Location and coverages of the various seismic reflection and bathymetric surveys are plotted in Figure 3.6.

In this study, we do not interpret the geologic structure and seismic-stratigraphy imaged on the BCS seismic reflection profiles. Rather, we focus on the morphology of the depth-converted structure map modeling the elevations of the basal erosional surface of Quaternary sediments derived from the interpretation of the seismic reflection data. In the BCS, that surface corresponds to the top Paleogene erosional surface. In this dissertation we refer to that surface as top pre-Quaternary erosional surface. The study of that surface is key to unravel the geomorphology and interrelationship of the Late Weichselian (and, perhaps, Late Saalian) Rhine–Meuse and Axial Channel palaeovalleys (see Mathys, 2009; De Clercq et al., 2016). Indeed, Holocene sediments lie directly on top of that erosional surface over most of the BCS and OTE areas (Figure 7.3). Hence, the top pre-Quaternary surface was possibly exposed to subaerial conditions, at least, during the Last Pleistocene marine lowstand.

The BCS top pre-Quaternary surface has been gridded at 25 m bin size to show areas with no data. We performed time–depth conversions by combining the seismic reflection data with a series of sediment cores collected along some of the seismic reflection profiles (see De Clercq et al., 2016). For data acquired outside the BCS, boreholes were not available. We therefore applied a constant sound velocity of 1560 ms^{-1}, which represents the average velocity estimated from the areas with available sedimentary information. The lack of boreholes to calibrate the time–depth conversion may have added some distortion to the top pre-Quaternary surface in those areas. However, this distortion should be negligible, as, apart from some isolated sandbanks, the sediment cover in the northwestern BCS and OTE is generally very thin.

We have also gathered several local depth-converted structure maps, exhibiting the top pre-Quaternary surface along some of the palaeovalleys carved into the OTE (See Chapter 3). These depth-converted structure maps were generated within the Osiris–Galloper project and have been provided for the present study as 11.7 m bin-size Surfer grids.

Detailed geomorphological analysis of the top pre-Quaternary surface already exists for the BCS area (e.g. Liu et al., 1993; Mathys et al., 2009; De Clercq et al., 2016). However, the geomorphology and recent evolution of the area where the Rhine–Meuse, Axial–Lobourg Channel and Thames palaeovalleys intersect one another remains poorly constrained. We will thus not describe the geomorphology of the top pre-Quaternary surface in much detail. Instead, we concentrate on major geomorphological features (e.g. streamlined scarps, etc.) possibly related to the continuation of the Rhine–Meuse palaeovalley(s) into the Axial Channel. With this, we aim to establish correlations between the morphology of the Axial–Lobourg Channel exposed at the seafloor and that sub-cropping on the BCS beneath the Holocene cover.

In this study, we interpret the geomorphology of the various palaeovalleys and palaeo-depressions by combining 2D–3D geophysical data interpretation (HIS Kingdom and Opendtect) with GIS analysis (Global Mapper and ArcMap).

7.4. Southern North Sea Geomorphology: Observations and interpretations

The seafloor of the southern North Sea exhibits several kilometer-scale sandbanks and dunes, deeply incised palaeovalleys, and apparently isolated palaeo-depressions (Figure 7.4). The formation of sandbanks and dunes is associated with the Holocene transgression and subsequent marine tides and submarine currents (e.g. Houbolt, 1968; Dyer and Huntley, 1999). The majority of the elongated, kilometer-scale sandbanks cluster in four main sandbank fields; i.e. the Thames Estuary Bank (TEB), Norfolk Bank (NB), the Flemish Bight Bank (FBB) and the Belgian–Dutch Continental Shelf Bank (BDCSB) (Figure 7.4; see also Houbolt, 1968; Dyer and Huntley, 1999). The Outer Thames Estuary, the Spur area and the Dover Strait also show some major elongated sandbanks (e.g. Dyer and Huntley, 1999). However, these are isolated from one another and do not form sandbank fields (Figure 7.4). Dunes are oriented obliquely to major sandbanks (e.g. Dyer and Huntley, 1999). The geomorphology and formation of the Holocene sandbanks and dunes are extensively described and discussed in the literature (e.g. Houbolt, 1968; Dyer and Huntley, 1999). We will therefore not discuss these features in the present study.

Sandbank fields do not extend over the western and southern parts of the Lobourg–Axial palaeovalley (Figure 7.4). Neither are there extensive sandbank or dune fields in the Dover Strait, nor in the OTE and in large parts of East Anglia. This distribution seems to be related to the direction and strength of Holocene tidal currents and the local palaeo-topography on which these sandbanks developed (e.g. Houbolt, 1968; Dyer and Huntley, 1999). Surprisingly, major sandbanks do not extend across a ~10 km wide belt separating the Norfolk Banks and Flemish Bight Banks (Figure 7.4). Similarly, the area between the Flemish Bight Banks and the Belgian–Dutch Continental Shelf Banks is not traversed by major sandbanks either. Importantly, the belt between the Norfolk Banks and Flemish Bight Banks without sandbanks is the area through which the outflow of the ice-marginal lake (or the southwestward-diverted German palaeo-river system) is inferred to have flowed during the Last Glacial Maximum (Figure 7.1; see Toucanne et al., 2010; Toucanne et al., 2015; Sejrup et al., 2016). The other area with no major sandbanks is the location where recent palaeogeographic reconstructions situate the Rhine–Meuse River system during part of the Saalian and the Weichselian glacial periods (Busschers et al., 2007; Hijma et al., 2012; Peeters et al., 2015, Peeters et al., 2016).

The most prominent palaeovalleys imprinted on the seafloor of the southern North Sea are the Lobourg and Axial Channels. In particular, the Lobourg Inner Channel (Lic) and the Axial Inner Channel (Aic) are very distinct in the bathymetry (see Figure 7.4). The similar orientation and general geomorphology of the Axial and Lobourg channels indicate that they are parts of the same NE–SW-oriented palaeovalley. Their inner channels (i.e. the Lic and Aic) are both carved along the western slope of the Lobourg–Axial Channel. The Lic and Aic are separated from each other at the seafloor by an isolated, ~25 m high, 56 km long, 2.6 km wide, NNE–SSW-oriented linear sandbank. Their alignment, morphology and similar NNE-trend suggest though that both inner channels are parts of the same channel.

Sandbanks and dunes cover large parts of the eastern Lobourg–Axial Channel, obscuring its eastern slope. Its western slope, on the other hand, is sharply marked in the seafloor, suggesting that this palaeovalley extends northward, at least until the area we labelled in Figure 7.4 as "North Axial Channel" (NAC). In there, a prominent N–S-oriented palaeo-channel (NAic) is sharply incised in the bathymetry. The western slope of that palaeo-channel is the continuation of the western slope of the Lic–Aic, suggesting that they are both part of the same palaeovalley system. The northward continuation of the NAic is unclear in the bathymetry, as the Norfolk Banks obscure its western edge. In addition, the relief of its eastern edge is much lower in the north than further south, making it difficult to distinguish it from Holocene sedimentary features (Figure 7.4).

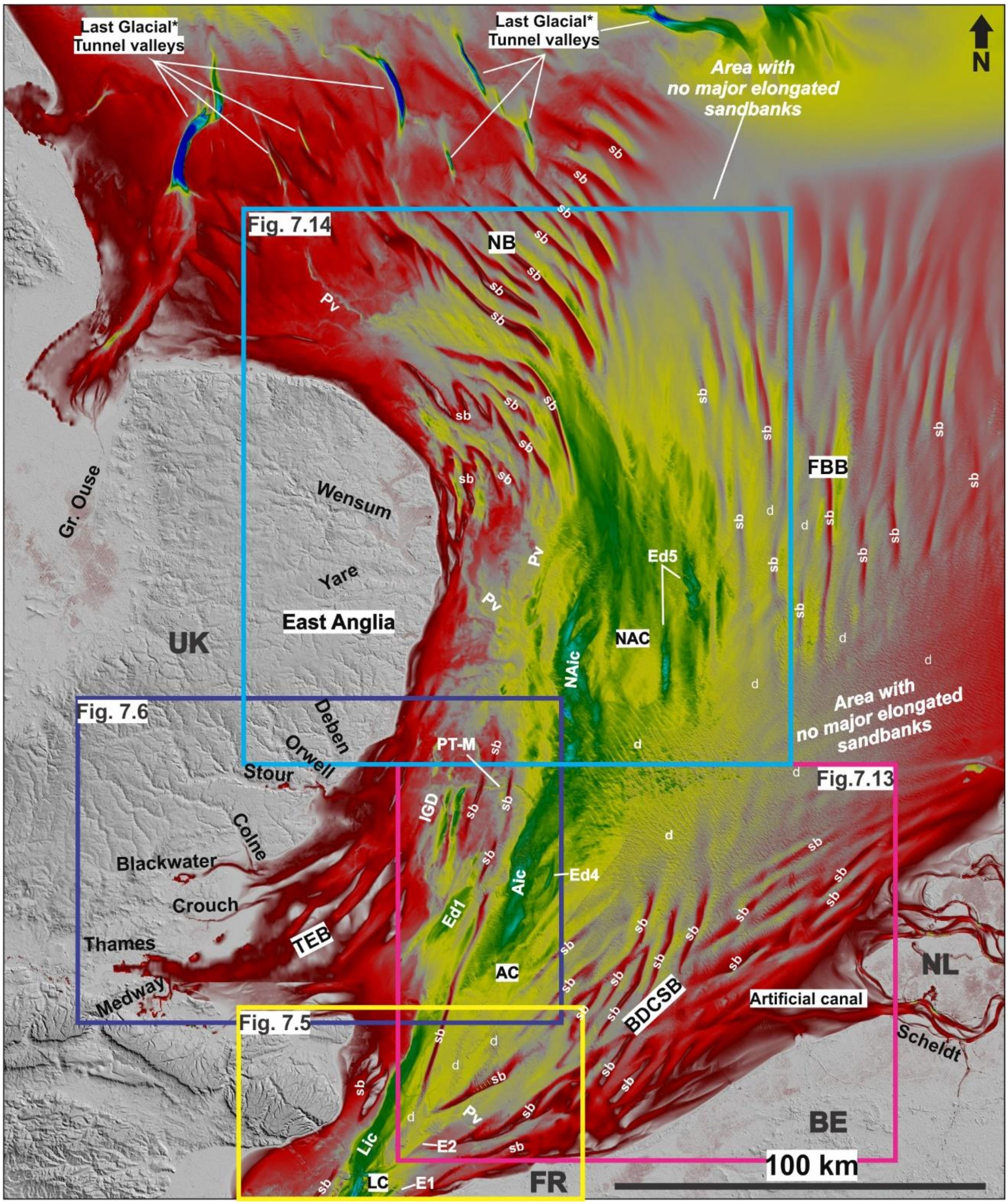

Figure 7.4. Bathymetric DTM resulted from the merging of the main bathymetric datasets used in this study (see Chapter 3). Major British Rivers are indicated. NB: Norfolk Banks; FBB: Flemish Bight Banks; BDCSB: Belgian–Dutch Continental Shelf Banks; TEB: Thames Estuary Bank; PT-M: Pre-Elsterian Thames–Medway (e.g. Bridgland and D'Olier, 1995); E1 and E2: Lobourg Channel Escarpments (see Chapter 6); Pv: palaeovalley; sb: Sandbank, d: dune and ripple fields; Ed1–5: elongated deeps; IGD: Inner Gabbard Deeps. For other labels see Figure 7.2; For Bathymetric color scale see Figure 7.5.

Other prominent smaller-scale palaeovalleys and palaeo-depressions are also imprinted on the seafloors of the OTE and northern Dover Strait areas, as well as in the Spur area and along the East Anglia offshore platform (Figure 7.4).

Here below, we provide detailed descriptions of the various geomorphological features mentioned above. We have divided the study area in 4 sectors (see Figure 7.4); these are: the northern Dover Strait, the Thames and outer Thames Estuary area, the southern Axial Channel and the North Axial Channel.

7.4.1. Geomorphology of the northern Dover Strait palaeovalleys

In this section, we extend further north the area discussed in Chapters 5 and 6 in order to complete the geomorphological characterization of the Lobourg Channel and to better show its relationship with other palaeovalleys. The geomorphology and possible formation of the Lobourg Channel is extensively described and discussed in Chapters 4, 5 and 6 (see also Collier et al., 2015). In the present Chapter, we will therefore not describe it again; rather, we summarize its major geomorphological characteristics.

The Lobourg Channel forms part of the Channel palaeovalley system (e.g. Collier et al., 2015). The geomorphology of this palaeovalley suggests the occurrence of three major erosional events. The first two carved escarpments E1 and E2 (Figure 7.5), which are remnants of early Lobourg Channel incisions; i.e. LC1 and LC2. The last (youngest) erosional event formed the Lobourg Inner Channel (Lic), which truncates the previous two. The erosional events that gave rise to the morphology of the present-day Lobourg Channel seem to have been rather intense, as this channel cuts through chalk and limestone (Figure 7.2). This is especially true for the Lic incision, which comprises several erosional features generally found in megaflood-eroded terrains. The Lic also truncates two palaeovalleys located on the British continental shelf, part of which are still expressed in the seafloor morphology as 500–650 m wide, 3–6 m deep hanging palaeovalleys (Figure 7.5b).

The morphology of the palaeovalleys flanked in the east by E1 and E2 is uncertain. Their absolute age is also unknown; they seem however to have been formed following the incision of the Fosses Dangeard (see Chapter 6). The morphology of the Lic is, on the other hand, sharply marked in the seafloor (Figure 7.5). The Lic is a NNE-trending, rectilinear, ~2.7 km wide palaeovalley characterized by box-shape cross-sectional morphologies and a flat bottom. This palaeovalley shows relief at the seafloor of 15–25 m across its western slope and 5–10 m across its eastern one.

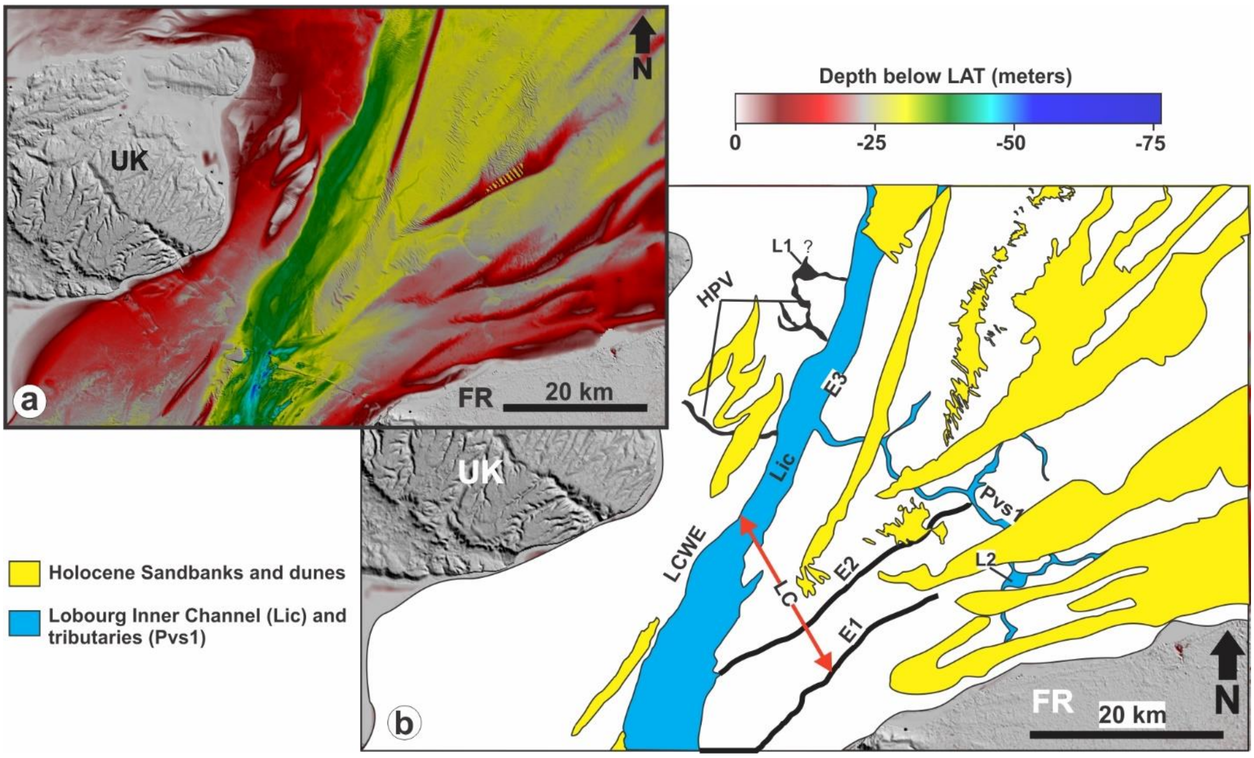

Figure 7.5. a) Bathymetric DTM of the Northern Dover Strait. b) Geomorphological interpretation of the bathymetry. HPV: Hanging palaeovalleys; LCWE: Lobourg Channel Western Edge; LC: Lobourg Channel; L1 and L2: palaeo-depressions within palaeovalley systems; E1, E2 and E3: Lobourg Channel Escarpments; Pvs1–9: palaeovalley systems (see next sections of this Chapter).

A complex tributary palaeovalley system (Pvs1) runs into the Lic (Figure 7.5b). Pvs1 is characterized by a series of palaeovalleys exhibiting widths of 0.6–1.2 km and reliefs at the seafloor of 2–3 m. This palaeovalley system comprises a ~3.5 km² depression (L2), which was most likely inundated when this palaeo-river system was active, forming a lake. Similar palaeo-depressions are also present along one of the hanging palaeovalleys identified in the British continental shelf (i.e. L1), and along several of the palaeovalleys systems identified in the Outer Thames Estuary (Figure 7.5 and Figure 7.6). The data available for this study are not enough to unravel how these depressions formed.

7.4.2. Geomorphology of palaeovalleys incised across the Thames and Outer Thames Estuary areas

The Thames Estuary (TE) and Outer Thames Estuary (OTE) areas exhibit several palaeovalleys and palaeo-depressions carved into substratum. Some of these are partially or entirely covered by a series of SW–NE-oriented, kilometer-scale, elongated sandbanks, the formation of which is related to the Holocene dynamics of the Thames Estuary (Figure 7.6a; see Dyer and Huntley, 1999). These Holocene sedimentary bodies do no extend across the OTE. There, there are some kilometer-scale, NNE–SSW-oriented, linear sandbanks that possibly formed during the Holocene marine transgression (e.g. Dyer and Huntley, 1999). However, these are scattered, covering only minor parts of partially infill and unfilled palaeovalleys and palaeo-depressions outcropping at the seafloor. Here below, we provide detailed descriptions of the main palaeovalley systems we have identified in the TE–OTE area, which we named Pvs2 – Pvs8. The interpreted geomorphology of the various palaeovalleys and palaeo-depressions is plotted in Figure 7.6c.

Pvs2 consists of a series of E–W-oriented minor palaeovalleys flowing out of or into a major N-S-trending palaeovalley (Figure 7.6). On the one hand, the E–W palaeovalleys are truncated either by Pv3 or by the western edge of the Lic–Aic, resulting in hanging-valley landforms. On average, these palaeovalleys show relief at the seafloor of 1.5–2.5 m and have widths of 400–800 m. On the other hand, the main N-S-trending palaeovalley is ~3 km wide and ~6 m deep. This palaeovalley seems to bifurcate into 2 channels in the south, one continuing toward the Lic–Aic and another extending southward. The latter evolves southward from a 0.5–1 km wide and 3–4 m deep palaeovalley to a NNE–SSW-elongated, 11–13 m deep, 1.5 km wide and 4 km long palaeo-depression (Ed2 (i) in Figure 7.6 and Figure 7.11). To the south, this depression is followed by another SSW-trending depression of similar depth (Ed2 (ii) in Figure 7.6). The geometry of Ed2 (ii) is unclear in the bathymetry, as a large sandbank covers part of it. A ~12 m high ridge separates depressions Ed2 (i) and Ed2 (ii). The composition of that ridge is unknown owing to the lack of sedimentary data from that location. This makes it impossible to assess the nature of the separation between these two depressions.

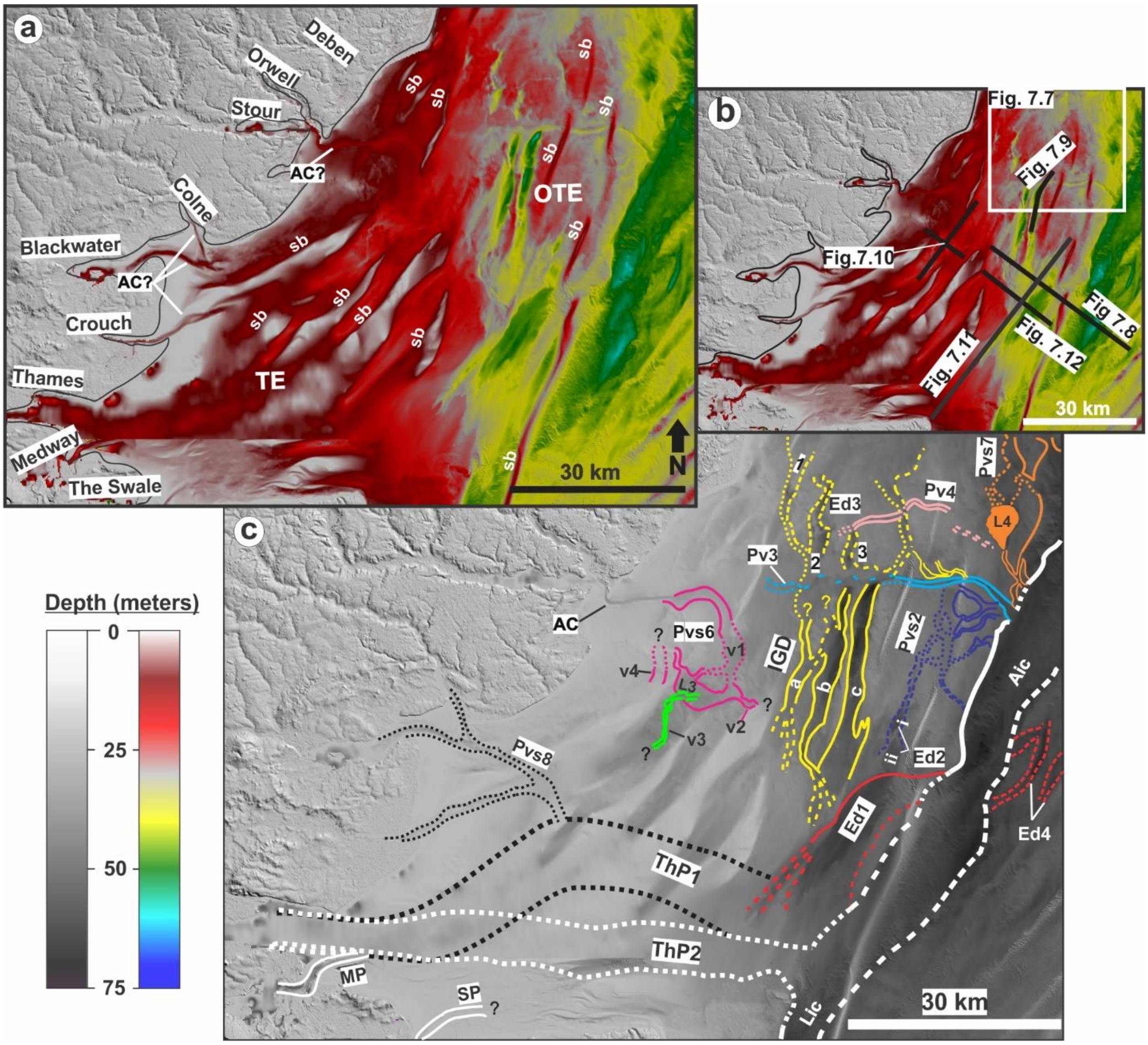

Figure 7.6. a) Bathymetric DTM of the Thames and Outer Thames Estuary. b) Location of profiles shown in Figure 7.8 to Figure 7.12 (black lines), and the area plotted in Figure 7.7 (white rectangle). c) Geomorphological interpretation of the bathymetry. MP: Medway palaeovalley; SP: Swale palaeovalley; AC: Artificial Canal; ThP: Inferred extents of Thames valley incisions. See Figure 7.2, Figure 7.4 and Figure 7.5 for other labels. For this and the following figures: continuous lines: features clearly imprinted on the bathymetry or observed in the seismic reflection data; dashed lines: inferred from the bathymetric and/or seismic reflection data; dotted lines: hypothesized based on our observations and/or previous studies; different colors: different ages.

Pv3 is an E–W-oriented, 0.7–1.2 km wide, and 5–6 m deep palaeovalley carved into substratum (Figure 7.7). This palaeovalley appears to have been a tributary of the Axial Channel. Its eastern half is sharply incised in the bathymetry, while its western half has been extensively eroded by the palaeo-depressions and palaeo-channels composing the system Ed3–IGD. The erosional process(es) that carved Ed3–IGD has removed almost completely the southern slope of Pv3. However, parts of its northern slope remains in the seafloor. In fact, the northern slope of Pv3 defines the northern head of the Inner Gabbard Deeps (Figure 7.9).

Pv3 was already identified by, among others, Bridgland and D'Olier (1995). These authors interpreted it as part of the pre-Elsterian Thames–Medway palaeo-river system. Contrarily to what these previous studies suggested, our bathymetric dataset shows that Pv3 does not extend to the British coastline. In fact, the palaeovalley apparently extending eastward from the Orwell–Stour River mouth (i.e. Pvs6) bends southward several kilometers to the west of the apparent western termination of Pv3.

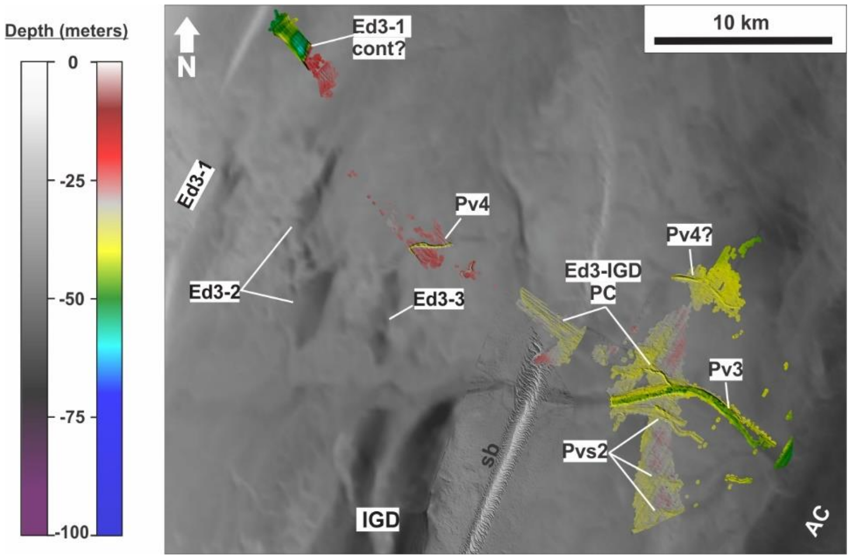

Figure 7.7. OTE top pre-Quaternary surfaces derived from seismic reflection data plotted on top of the bathymetric DTM. Note that palaeovalleys and palaeo-depressions identified in the OTE are carved into Paleogene substratum. "Ed3-1 cont" indicates a buried palaeo-depression possibly linked to Ed3-1.

Pv4 is a 500–800 m wide, 1–2 m deep palaeovalley carved into substratum, extending eastward parallel to Pv3 (Figure 7.7). Neither the deeps composing the Ed3–IGD system nor its associated palaeo-channels seem to truncate this palaeovalley, suggesting that Pv4 was formed following the incision of that system. However, the bathymetric dataset is not accurate enough at the intersection of these features to corroborate that.

The system Ed3–IGD evolves toward the south from a series of NNE–SSW-oriented palaeovalleys and palaeo-depressions to three major NNE–SSW-elongated deeps; i.e. The Inner Gabbard Deeps (Figure 7.6; Emu Ltd & University of Southampton, 2009). This system can be divided into two groups according to whether they are carved to the north or to the south of Pv3. These are respectively: Ed3 and The Inner Gabbard Deeps (IGD). These palaeo-depressions are not continuous across Pv3. They appear however to be connected across that palaeovalley by several minor palaeo-channels.

Ed3 consists of 3 incisions labeled 1, 2 and 3 in Figure 7.6. Palaeo-depression Ed3-1 incises up to 35 m into substratum, although only 13 m of its depth is exposed at the seafloor (Figure 7.7). Its width is not clear in either the bathymetry or the seismic reflection data. Palaeo-depressions Ed3-2 and Ed3-3, on the other hand, appear to have thin to no sedimentary infill (Emu Ltd & University of Southampton, 2009). However, it is possible that part of these palaeo-depressions are buried, as the seismic reflection profiles available from that area are too sparse to image their three-dimensional geometry. Ed3-2 consists of two NNE-elongated depressions. The northern palaeo-depression is 1–2 km wide and shows reliefs at the seafloor of 15–20 m, while the southern depression exhibits widths of 2–2.7 km and incises 10–20 m into substratum. Finally, the exposed part of Ed3-3 comprises one 1–2 km wide, ~10 m deep palaeo-depression.

The IGD extend southward over 20–23 km from about the northern edge of Pv3 (Figure 7.6). This feature comprises three major elongated interconnected palaeo-depressions carved into substratum, in this paper referred to as depressions “a”, “b” and “c”. Depression “a” is ~20 km long, ~2 km wide and 5–8 m deep (Figure 7.8). Depressions “b” and “c” are separated from each other by a prominent 1–2 km wide ridge, which rises 20–25 m above their bottoms (Figure 7.6). This ridge widens and flattens southwestward. Depression “b” is ~22 km long, 1.6–3 km wide and 25–30 m deep. This depression evolves southward into a southeast-trending, ~800 m wide, 8–10 m deep, partially infilled palaeovalley (Figure 7.12). The eastern end of this palaeovalley forms a hanging-palaeovalley landform at its intersection with palaeo-depression Ed1 (Figure 7.6). Depression “c” extends over ~31 km, exhibiting depths at the seafloor of up to ~28 m. Contrarily to depression “b”, this depression widens to the south, from 2.5 km to 4.7 km. However, both depression “b” and “c” show similar irregular U-shaped cross-sectional morphologies (Figure 7.8). They both also shallow from their northern heads toward the south. Importantly, depression “c” exhibits a SSW-trending, scoop-shaped, overdeepened area at the base of its northern slope (Figure 7.9), which may attest to fluvial scouring (i.e. plunge-pool erosion). Indeed, there is a drop of ~20 m from the northern edge of Pv3 to the bottom of depression “c”, which would have formed a waterfall anytime runoff flowing southward passed this way, possibly inducing plunge-pool erosion at its base.

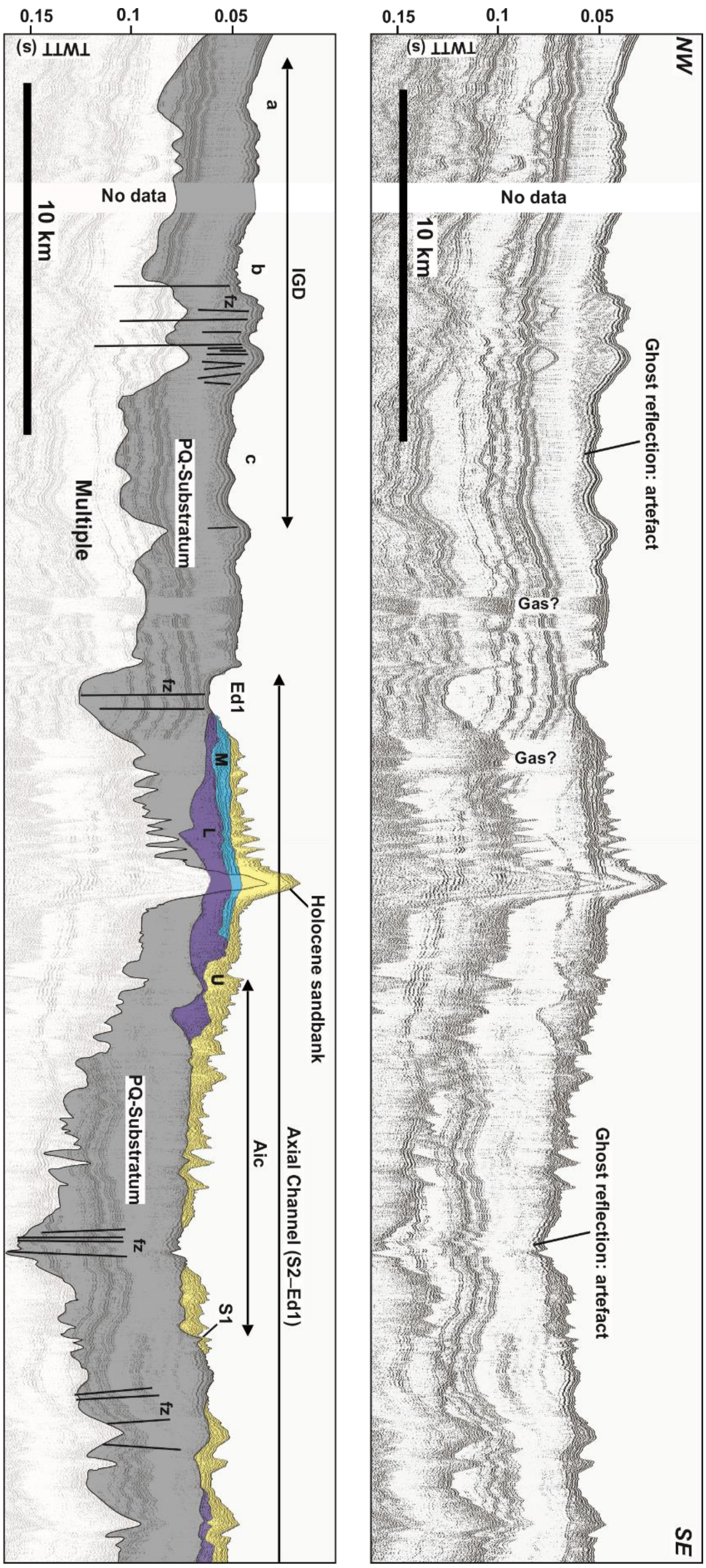

Figure 7.8. Seismic reflection profile traversing Aic, Ed1 and IGD, non-interpreted (above) and interpreted (below). PQ-Substratum: pre-Quaternary substratum; U: Upper Unit; M: Middle Unit; L: Lower Unit; Multiple: seismic multiples (i.e. Artefacts); fz: fault zones typical of some of the clay units composing the southern North Sea Paleogene and Neogene bedrock (see also Figure 7.11; Le Bot et al., 2003) – this definition also applies to features labelled "fz" in other figures.

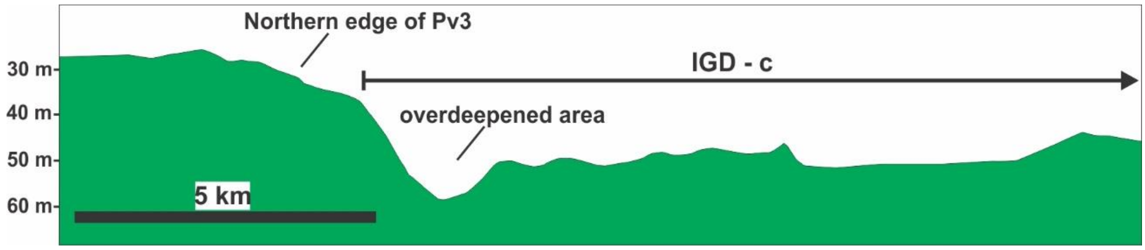

Figure 7.9. Topographic profile across the length of depression "c" of the IGD.

Pvs6 consists of a series of palaeovalleys running into or outflowing from a ~20 km² depression (L3), which possibly formed a lake when these palaeo-rivers were active (Figure 7.6 and Figure 7.10). The main palaeovalley (v1) of Pvs6 extends over several kilometers eastward, possibly from about the mouth of Rivers Orwell and Stour, before bending clockwise toward L3. This palaeovalley shows depths at the seafloor of 1–3 m and widths of 1–2 km. Its southern continuation is not very distinct in the bathymetry. In fact, apart from v3, the various palaeovalleys composing Pvs6 are not well defined in the bathymetry (Figure 7.6). Palaeovalley v3, on the other hand, is sharply incised in the seafloor, exhibiting N–S orientations, widths of 500–750 m and depths at the seafloor of 3–5 m. The part of this palaeovalley visible in the bathymetry extends over ~12 km, trending toward Pvs8. Most palaeovalleys and palaeo-depressions composing Pvs6 are partially infilled with sediments (Figure 7.10). Exceptionally, v3 truncates the infills of L3 and that of a buried palaeovalley (Figure 7.10). Palaeovalley v3 thus appears to represent the last phase of valley incision along Pvs6. The lack of a denser seismic reflection grid in that area prevents to characterize the geometry and extent of other palaeovalleys of this system, as well as to confidently establish the relationship (i.e. relative age) between IGD and Pvs6, and between Pvs6 and Pvs8.

Pvs8 has been dredged near shore. The system of canals observed in the bathymetry has therefore been carved artificially. The morphology and extent of Pvs8 shown in Figure 7.6 comes from maps shown in Bridgland and D'Olier (1995). These authors proposed that these buried palaeovalleys represent tributaries of the Late Saalian and/or Weichselian Thames Palaeo-river.

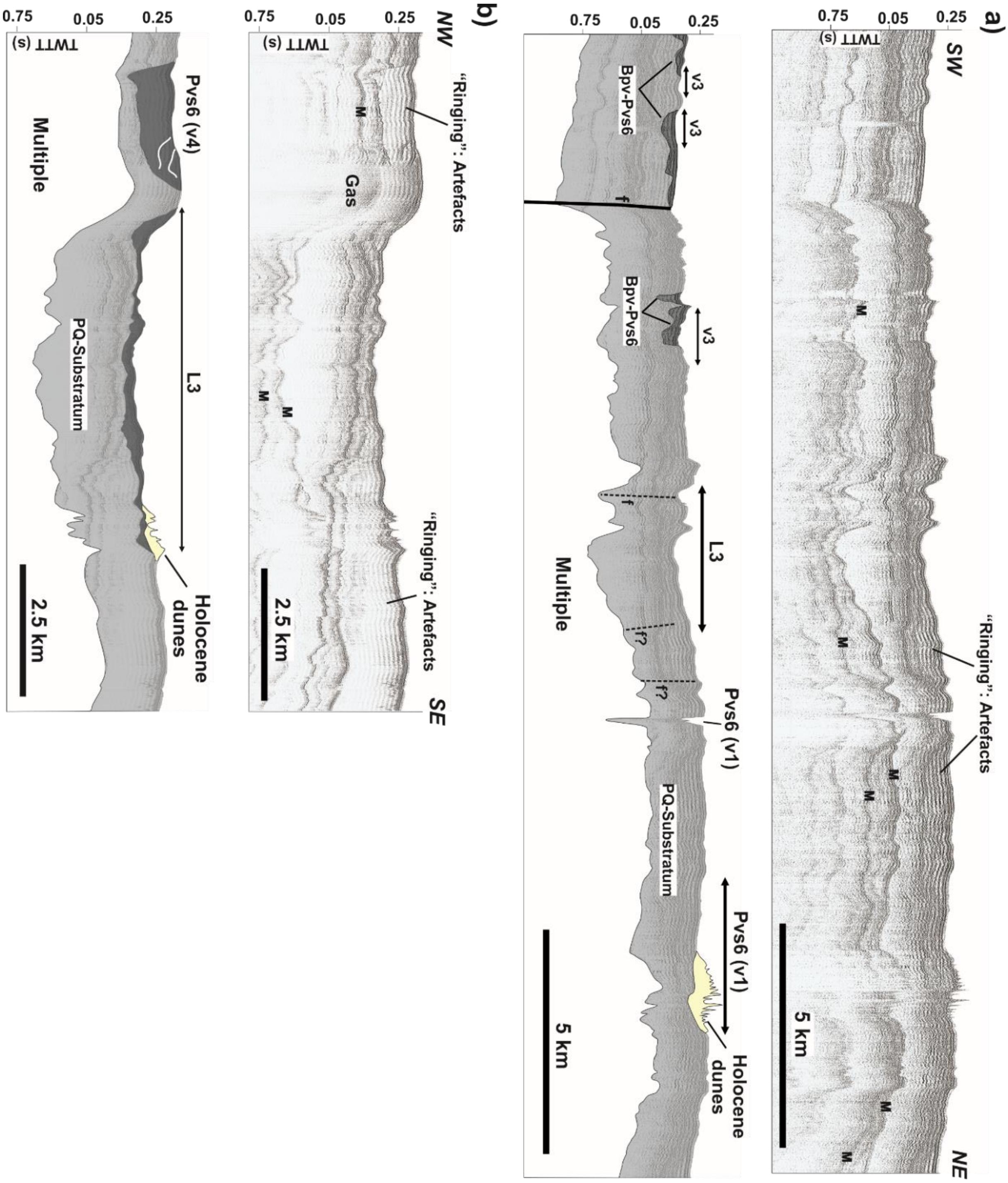

Figure 7.10. Seismic reflection profiles traversing Pvs6 and L3, non-interpreted (above) and interpreted (below). Black transparent areas in (a): Pvs6 buried palaeovalleys (Bpv); Black transparent areas in (b): infill of palaeo-depression L3 and v4 of Pvs6. v3: palaeovalley v3 of Pvs6 observed in the bathymetry; M: multiple reflections (i.e. artefacts).

The morphology of palaeovalleys extending eastward from the Thames mouth is obscured in the seafloor by Holocene estuary sediments deposited on top (Figure 7.6a). The buried Thames palaeovalleys imaged in our seismic reflection dataset exhibit two distinct seismic units (Figure 7.11). These appear to be the infill of two channels, one (ThP1) located northeastward from the other (ThP2). Palaeovalley ThP2 appears to truncate ThP1, indicating that they formed during different erosional episodes. Both palaeovalleys ThP1 and ThP2 are truncated by a rather flat erosional surface covered with apparently prograding reflections. Its seismic facies significantly resembles the seismic facies interpreted as sediments deposited on top of the Holocene marine transgression identified in the BCS (Mathys, 2009). It is thus possible that that erosional surface represents the Holocene marine transgression.

ThP1 shows depths of up to 16–17 m (calculated assuming sound velocities of ~1600 ms^{-1}) and widths of 9–10 km. The infill of this palaeovalley is characterized by acoustically transparent seismic facies. ThP2, on the other hand, is ~17 m deep and ~7 km wide; its infill exhibits high-amplitude cross-bedded reflections (Figure 7.11). The infill of ThP2 is interrupted by two prominent internal erosional surfaces, indicating the occurrence of several infilling-and-erosional episodes. The infill of ThP1 appears to be more homogeneous, exhibiting no visible internal erosion. ThP1's infill is actually rather similar to the Lower Unit infilling the Axial Channel (Figure 7.8; see next section of this Chapter). They may thus be remnants of the same infilling episode. We have tentatively mapped the Thames palaeovalleys (ThP1 and ThP2) in Figure 7.6 by combining our observations with maps published in previous studies (e.g. Bridgland and D'Olier, 1995).

Ed1 is a NNE–SSW-oriented, ~6 km wide, ~18 m deep, elongated depression, which extends obliquely to the IGD and runs parallel to the Lobourg–Axial Channel (Figure 7.6). This depression cuts through substratum and truncates the lower seismic unit infilling the Axial Channel and the sediments infilling the Thames palaeovalley "ThP1" (Figure 7.8, Figure 7.11 and Figure 7.12). The incision of Ed1 thus postdates the formation and infilling of those palaeovalleys.

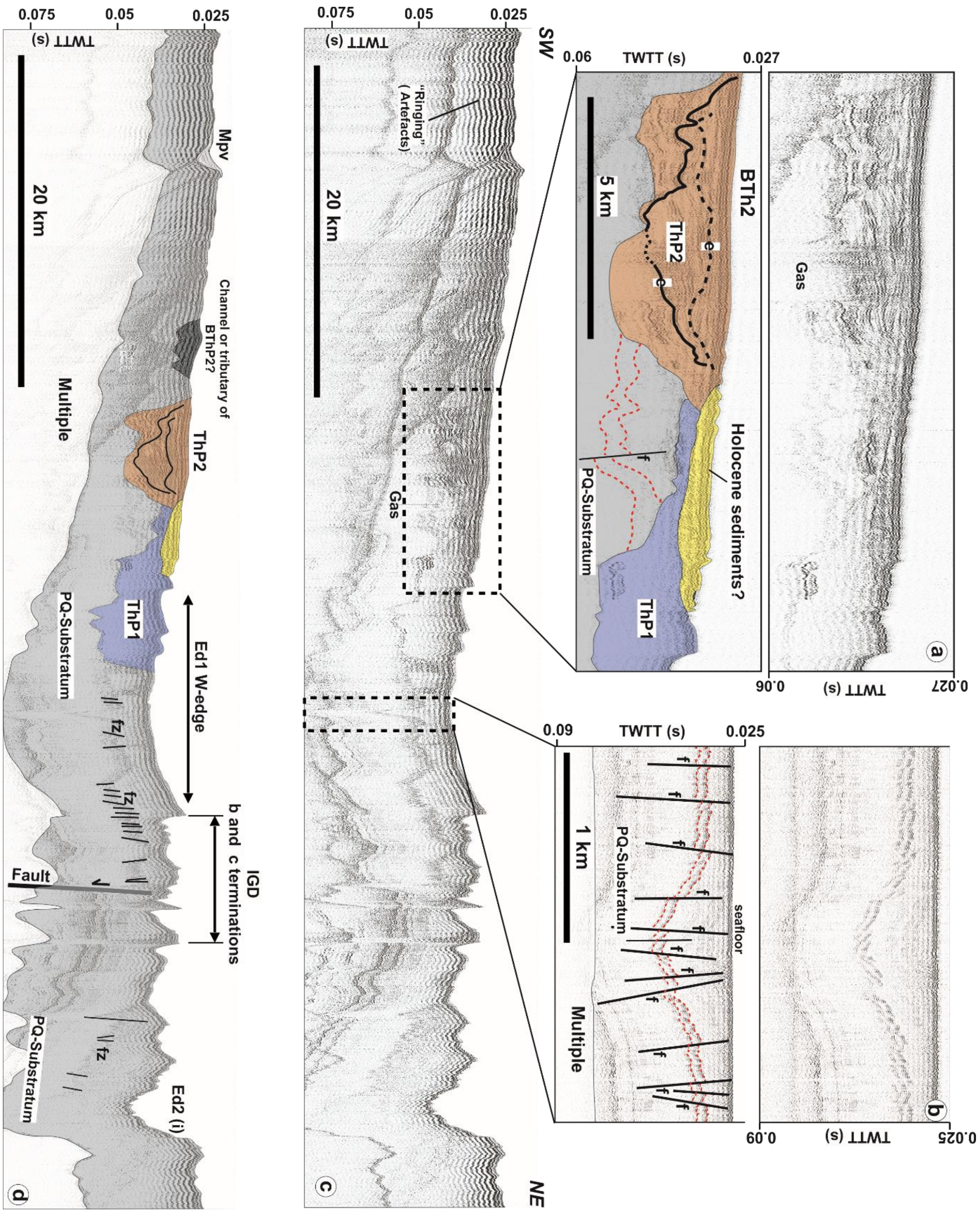

Figure 7.11. Seismic reflection profile acquired across Ed2, IGD and the Thames palaeovalleys, and along the western edge of Ed1. a) Enlarged area showing the Thames palaeovalleys non-interpreted (above) and interpreted (below). b) Enlarged area exhibiting faulting in clay units demonstrating that those reflections belong to Paleogene/Neogene bedrock. c) Complete seismic reflection profile non-interpreted and interpreted (d). Mpv: Minor buried palaeovalleys; e: internal erosional surfaces, f: faults; Red dashed lines: selected seismic horizons showing geometry of substratum's strata.

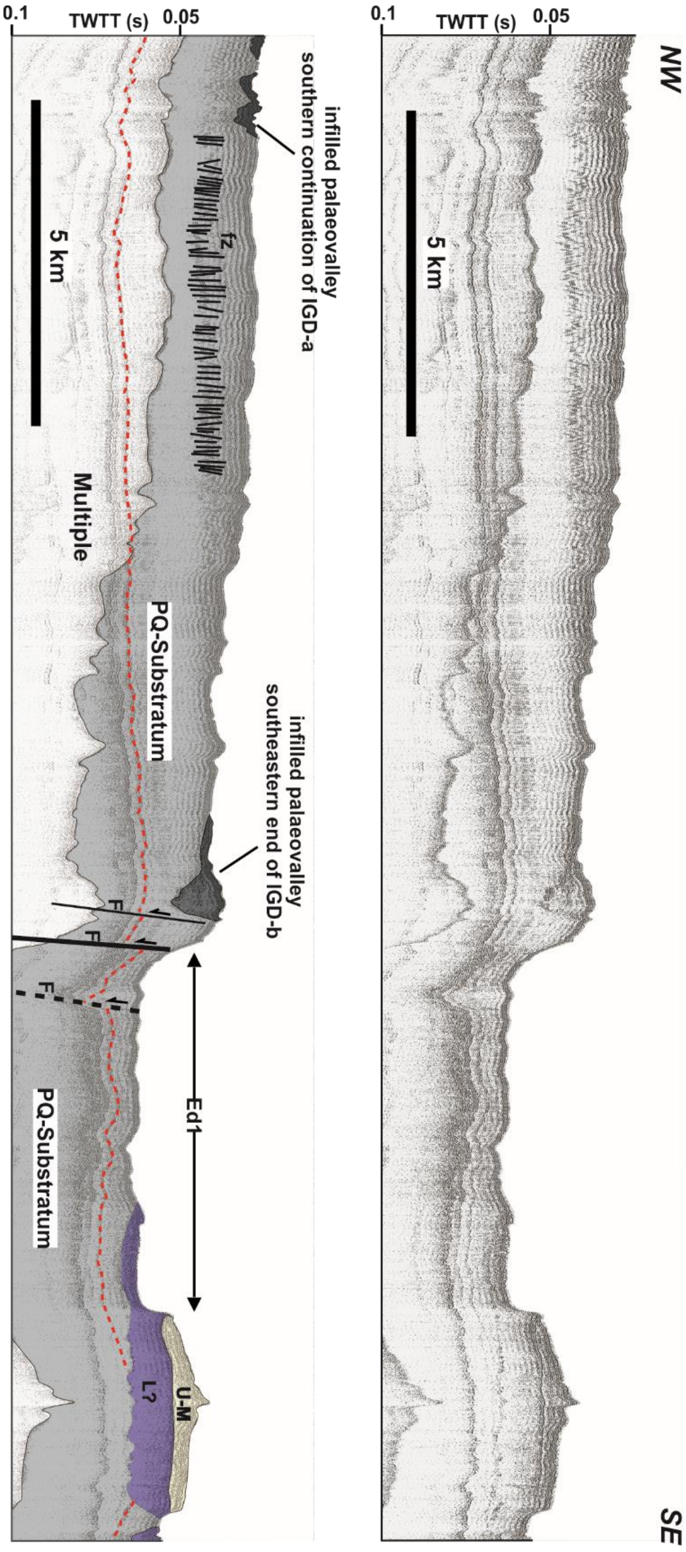

Figure 7.12. Seismic reflection profile acquired across Ed1 and the southern terminations of depressions "a" and "b" of the IGD. U–M: Possible Upper and/or Middle units infilling the Axial Channel; L: Lower Unit infilling the Axial Channel; PQ-Substratum: pre-Quaternary substratum. Red dashed line: seismic horizon showing geometry of substratum's strata.

7.4.3. Geomorphology of the southern Axial Channel

The geomorphology of the entire Axial–Lobourg Channel is not distinct in the bathymetry due to extensive sandbank and dune fields covering most of its southeastern part (Figure 7.4). Neither does the bathymetric data show much information on the connections between this palaeovalley with the Rhine–Meuse and Thames palaeovalleys. The northward continuation of the Lobourg Inner Channel (Lic) is also partially obscured by Holocene sediments. However, as already mentioned, the bathymetry suggests that this channel continues northward as the 8–10 km wide, 10–15 m deep Axial Inner Channel (Aic).

The seismic reflection data shows a series of NE–SW-oriented scarps carved into the top pre-Quaternary surface across the BCS (i.e. S1–S6), which may be linked to the Middle–Late Pleistocene evolution of the southwestern part of the Axial Channel. The most prominent of these scarps are S1, S2, S3 and S4 (Figure 7.13).

S1 is a 10 m high scarp, which defines the eastern edge of the Aic (Figure 7.13). Seismic reflection profiles across this channel show that it is partially incised into sediments infilling the Axial Channel. Note that Aic has never been identified prior this study.

S2 and S3 run sub-parallel to each other and obliquely to S1 (Figure 7.13). S2 appears to be 20 m high at its intersection with S3, although this may be due to local overdeepening. Indeed, S2 appears to truncate S3, which would add the height of S3 to the latter in that location. In addition, these scarps meet at the mouth of a significant tributary (T in Figure 7.13). Elsewhere, S2 and S3 are ~10 high and 4–10 m high, respectively. Parts of S2 and S3 were already identified by Liu et al. (1992), who interpreted S2 as the southwestern slope of the Axial Channel. However, these authors did not associate this scarp with any specific age. More recently, Hijma et al. (2012) modeled the 30–25 ka Rhine–Meuse–Axial palaeovalley assuming scarp S3 as its southeastern slope (see Figure 7.1). Hence, these two scarps may represent two phases of valley incision along the Late Weichselian Rhine–Meuse–Axial palaeovalley.

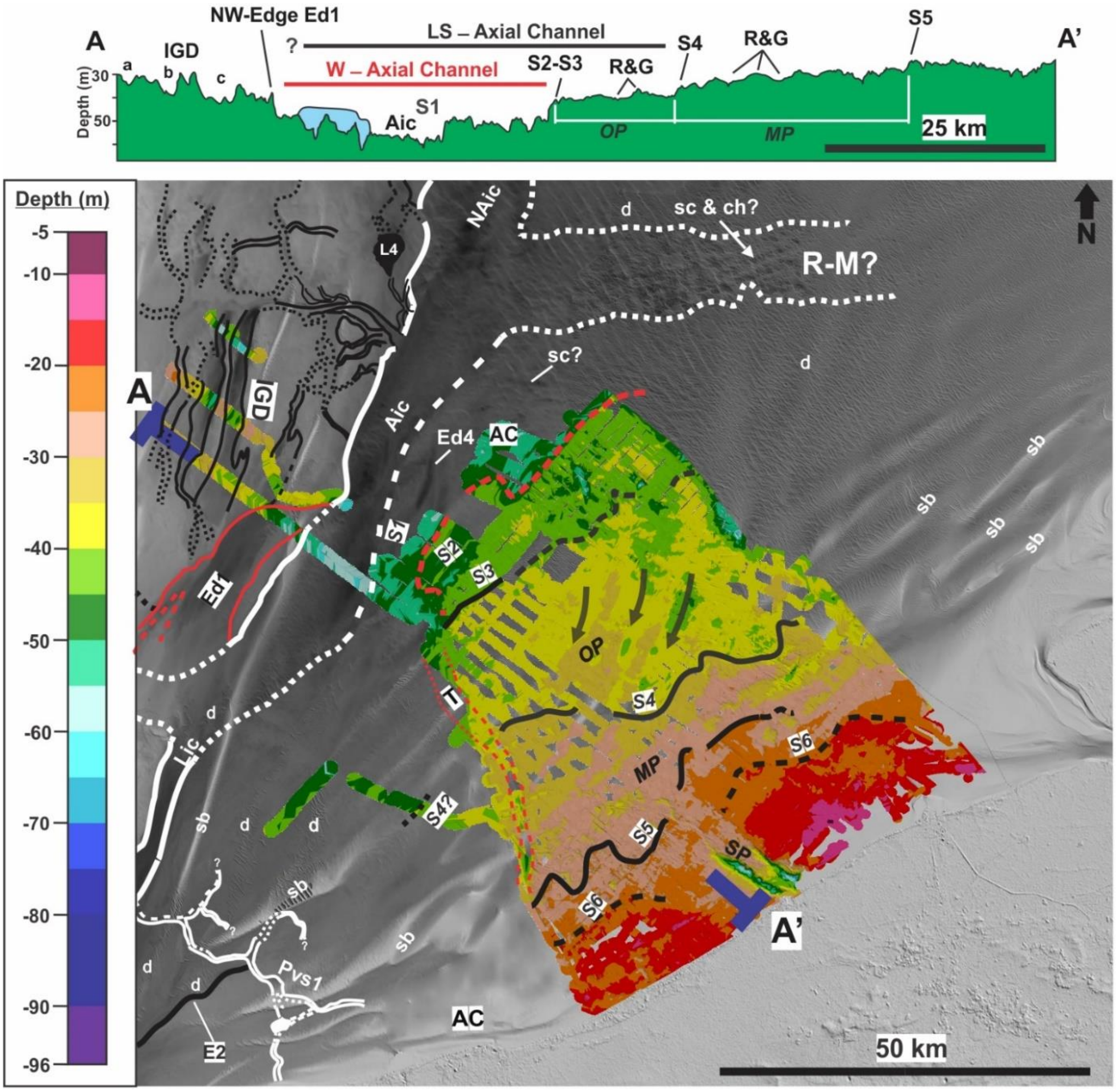

Figure 7.13. Below, BCS top pre-Quaternary surface plotted on top of the bathymetry. Above, A–A' cross-section (extremities marked in blue in the map) showing the topography of the top pre-Quaternary surface across the BCS, Axial Channel and IGD. Light Blue colored area in A–A': Lower–Middle seismic unit infilling part of the Axial Channel (see Figure 7.8); LS–Axial Channel: possible width of the Late Saalian Axial Channel; W–Axial Channel: Width of the Weichselian Axial Channel defined by S2/S3–Ed1; sc & ch: possible partially buried scours and/or channels; MP: Middle Platform; OP: Offshore Platform; SP: Sepia Pits; white lines: last phase of valley incision (Lic–Aic); Red lines: features possibly linked with S2–Ed1 valley incision; Black lines: scarps possibly older than S2; R-M: Rhine–Meuse palaeovalley. Arrows indicate orientations of grooves carved into the offshore Platform. For others labels see Figure 7.2, Figure 7.4 and Figure 7.6.

It is unclear from the available data to which phase of incision (S2 or S3) the tributary palaeovalley intersecting these scarps corresponds. This palaeovalley was already identified by previous studies (Mathys, 2009) and interpreted by Hijma et al. (2012) as the 30–25 ka Scheldt palaeovalley (see Figure 7.1). However, we have not been able to corroborate that, as its southeastern continuation is not distinct in the BCS top pre-Quaternary surface.

The infill of the Axial Channel exhibits three distinct seismic units in the seismic reflection data (Figure 7.8). The Lower Unit is characterized by acoustically almost-transparent seismic facies. The Middle Unit consists of sub-horizontal high-amplitude reflections. Finally, the Upper Unit is characterized by acoustically transparent and chaotic seismic reflections. According to BGS geological maps (see Figure 7.3), the areas where we have identified the Lower and Middle seismic units match locations infilled with Late–Middle Pleistocene sediments. However, the lack of sediment cores from those areas prevents a more accurate characterization of these units. The Upper Unit, on the other hand, shows similar seismic facies to those interpreted in other parts of the BCS as Holocene sediments (e.g. Houbolt et al., 1968; Mathys, 2009).

Palaeo-depression Ed1 and palaeo-channel Aic are partially carved into the Middle and Lower Units infilling the Axial Channel (Figure 7.8 and Figure 7.12), attesting to the younger ages of those erosional features. Based on the similar relative heights of Ed1 and the intersection of scarps S2 and S3, and the fluvial valley-like cross-sectional morphology exhibited by profiles across S2 and Ed1 (Figure 7.13), we postulate that Ed1 was carved during S3 and/or S2 incisions. We therefore interpret S2 and the western edge of Ed1 as slopes of a major palaeovalley, which represents the penultimate phase of valley incision along the Axial Channel observed in our dataset. The cross-sectional morphology defined by Ed1–S2 suggest that the Aic was carved into the palaeo-river bed of that palaeovalley. The Aic and its southern continuation (i.e. the Lic) thus seem to represent the last significant phase of valley incision that occurred along the Axial Channel. The younger age of the Aic is corroborated by the fact that the deeps Ed4, which are carved within the Axial Channel defined by S2 and Ed1, and possibly formed during that or subsequent valleys incisions, are truncated by the Aic (Figure 7.6 and Figure 7.13).

Other scarps imprinted on the BCS top pre-Quaternary surface are generally not higher than 4–5 m (Figure 7.13a), although some of them show relief up to ~10 m locally. These scarps (S4, S5 and S6) have already been identified in previous studies (e.g. Liu et al., 1992; Mathys, 2009; De Clercq et al., 2016). Mathys (2009) postulated that S4, which is typically referred to as the “Offshore Scarp” (e.g. De Clercq et al., 2016), may represent the southeastern slope of the palaeo-river system formed by the Rhine and Meuse palaeo-rivers during Late Saalian glacial maximum (see also Hijma et al., 2012). Finally, S5 and S6 are associated with possible

Pleistocene marine transgressions (e.g. Liu et al., 1992).

The combination of the top pre-Quaternary depth-converted structure map with the bathymetry shows that S1, S2 and, possibly, S3 meet to the northeast with a series of partially buried E–W-oriented features resembling scours and/or palaeo-channels (Figure 7.13). These features do not seem to be truncated by any of the scarps identified in the BCS. That implies that scarps S1, S2, S3 and S4 either are truncated by these features or bend eastward toward the position from where the Rhine–Meuse palaeo-river entered the southern North Sea during Late Saalian and Weichselian Periods (Figure 7.13). This suggest that these scarps may represents different phases of valley incision along the western continuation of the Rhine–Meuse palaeo-river's southern slope. Hence, this supports previous models suggesting that the Axial Channel represents the western continuation of the Rhine–Meuse palaeovalley during Late Saalian and Weichselian glacial maxima (Figure 7.1; e.g. Liu et al., 1992; Hijma et al., 2012).

Even though dunes obscure most of their morphology, many of the W–E-oriented channels and scours possibly associated with the Rhine–Meuse palaeovalley seem to end at the Aic's eastern edge (Figure 7.13). However, none of the previous studies modeling the extent of the various phases of evolution of the Rhine–Meuse palaeo-river since the Late Saalian glacial maximum extended this palaeovalley that far west (Figure 7.1). This suggests that, at some point, the palaeo-river flowing along the Axial Channel concentrated in the Lic–Aic. This palaeovalley thus became the main river, into which the Rhine–Meuse and other palaeo-rivers (e.g. Thames, Pvs1, etc.) ran.

Other remarkable geomorphological features imprinted on the BCS top pre-Quaternary surface are a series of NE–SW-oriented ridges and grooves and two buried palaeo-depressions known as the Sepia Pits (SP in Figure 7.13). The ridge-and-groove bed-forms are carved in the Offshore Platform (OP in Figure 7.13), showing reliefs up to 7–10 m. Some of them appear to cut through S3, suggesting that the erosional event(s) that carved these features postdate(s) the formation of that scarp. The Sepia Pits consist of two, SE–NW-oriented, elongated depressions, extending over ~10 km (see also De Clercq et al., 2016). These depressions exhibit depths at the top pre-Quaternary surface of ~30 m and widths of 1.5–2.3 km. They appear to be incised within a palaeovalley, known as the Oostende Valley (e.g. Liu et al., 1992). The formation of the Sepia Pits/Oostende Valley is associated with Late Saalian fluvial erosion followed by overdeepening due to posterior (Eemian) marine erosion (e.g. Liu et al., 1992; Mathys, 2009).

7.4.4. Geomorphology of the North Axial Channel

The North Axial Channel (NAC) presents a series of significant erosional features, including N–S-trending palaeo-channels and palaeo-depressions (Figure 7.14). The most prominent of these are: a deeply incised N–S-oriented inner palaeo-channel (NAic) and a NE–SW-oriented low-relief area located between the Flemish and Norfolk Banks with no sandbank development.

The NAic is a 12 km wide, ~20 m deep, N–S-oriented palaeovalley (Figure 7.14). It exhibits a 7–8 m high, ~9 km long, elongated promontory carved along its channel bed. This promontory shows similar morphology to palaeo-streamlined islands identified in the Dover Strait (see Chapters 5 and 6) and in the English Channel (Collier et al., 2015), which have been associated with megaflood erosion. NAic is sharply marked in the bathymetry over ~40 km (Figure 7.14). Its western slope is actually the continuation of the Aic. Therefore, the NAic and the Lic–Aic channel appear to be parts of the same inner channel and, so, formed during the same erosional episode.

The northern continuation of the NAic is unclear in the bathymetric data. Its western edge is completely covered by dunes and sandbanks, obscuring its geometry. Its eastern edge, on the other hand, enters into a rather flat area. This slope appears to continue northward as an N–S-oriented, 4–5 m high scarp (Figure 7.14). However, the absence of seismic reflection data in that area prevents to evaluate whether that scarp is associated with the NAic incision or with posterior sedimentary/erosional processes.

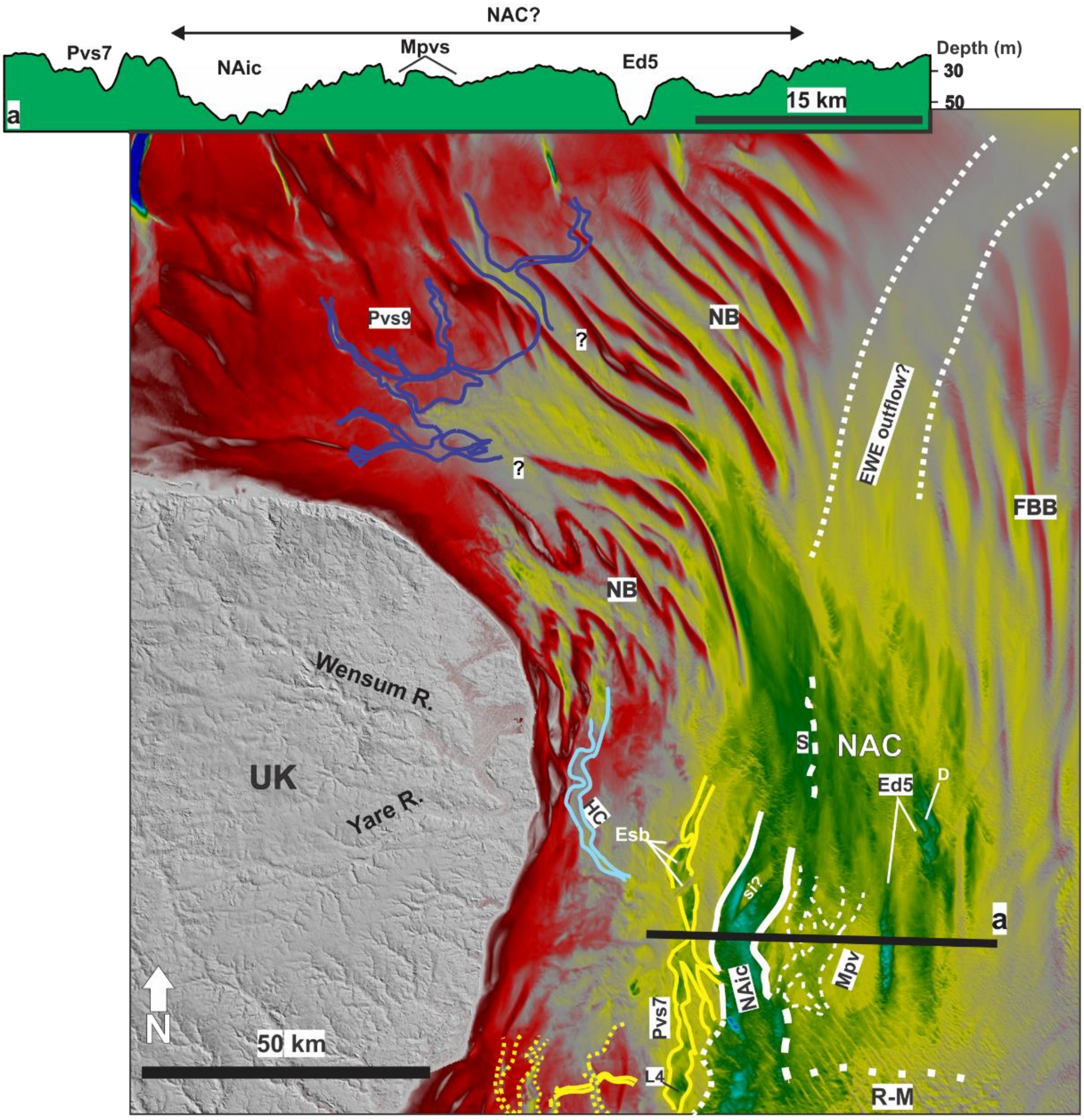

Figure 7.14. Bathymetric DTM of offshore East Anglia, the Flemish Bight and Spur areas. (a) Topographic cross section across Pvs7, NAic, Mpv and Ed5. Bathymetric color scale is shown in Figure 7.5. Main geomorphological features are indicated. HC: Holocene Channel; Esb: Possible Eemian sandbanks; D in Ed5: possible Holocene dunes; S: possible northward continuation of NAic eastern edge; EWE outflow: hypothesized outflow of a proglacial lake or channel formed by Weichselian southward diverted Ems, Weser, and Elbe Rivers. Yellow lines: palaeovalleys older than EWE outflow; Blue lines: palaeovalleys younger than EWE outflow (i.e. Pvs9 and HC). See previous figures for other labels.

Northeastward of the NAic, a 10–15 km wide, NE–SW-oriented low-relief area extends in between the Norfolk and Flemish Bight sandbank fields (Figure 7.14). Sandbanks are not continuous through that area (Figure 7.4). Importantly, the area with no sandbank development coincides with the outflow hypothesized by Sejrup et al. (2016) for a Late Weichselian (30–19 ka BP) proglacial lake, as well as with the path theorized by Toucanne et al. (2010; 2015) for the river formed 30–18 ka BP by the confluence of southward diverted northwestern German rivers (Figure 7.1). It is therefore possible that the disconnection between sandbanks across this belt might be due to the presence of a major buried palaeovalley in that area. That is, if a palaeovalley was incised there before the Holocene transgression, Holocene sediments would have had to infill the palaeovalley first before the sandbanks developing at its sides could extend across it. The presence of buried palaeovalley(s) may also explain why sandbanks from the Flemish Bight and the Belgian–Dutch Continental shelf do not connect to one another either (Figure 7.4). Indeed, the area without major sandbanks between those sandbank fields coincides with the possible location of the Weichselian Rhine–Meuse palaeovalleys (Figure 7.1; e.g. Busschers et al., 2007; Hijma et al., 2012). Of course, for the moment, this is just a hypothesis, as no seismic reflection data were available from those areas to corroborate it.

Other prominent palaeovalleys and palaeo-depressions incised in the North Axial Channel area are (Figure 7.14): Ed5, Pvs7, Mpvs, Pvs9 and a NNW–SSE-oriented Holocene palaeovalley (HC). We describe their geomorphologic characteristics here below.

Depressions Ed5 exhibit widths of 3.5–4.5 km, lengths of 20–25 km and depths at the seafloor up to 18 m. These elongate deeps are carved outside the Flemish Banks, which indicate that they are not troughs between parallel sandbanks. They also present some dunes at their bottoms, extending obliquely to the orientation of their main axis, suggesting that they were not formed during the Holocene. The various depressions composing Ed5 are subparallel to and have similar depths to the NAic (Figure 7.14).

Pvs7 extends parallel to the NAic over ~80 km, cutting through pre-Elsterian Quaternary substratum (Figure 7.2). Its northern half is characterized by a single channel, exhibiting widths of 2–5 km and depths at the seafloor of up to 8–12 m. This palaeovalley is traversed across its width by two prominent ridges (Figure 7.14). These ridges were interpreted by Limpenny et al. (2011) as remnants of Eemian sandbanks. If that is so, Pvs7, or at least its northern half, has not hosted large rivers since Late Saalian times (MIS 6), as otherwise, those sand ridges would have been washed out. Its southern half, on the other hand, evolves toward the south into three main channels, all apparently running into the NAic. Neither of these seem to be truncated by the NAic; they look like contemporary tributaries (Figure 7.14). The easternmost channel of the southern Pvs7 system widens southward, ending in a ~7 km long and 3 km wide depression (L4), which is carved up to 7–8 m into Pvs7's bottom.

Mpvs refers to a series of 6–8 m deep, 1.5–2.5 km wide, south-trending palaeovalleys that extend parallel to the eastern edge of NAic (Figure 7.14). These palaeovalleys seem to be truncated by the NAic, suggesting that they were formed during previous erosional phases.

Pvs9 comprises a series of NW–SE and NNW–SSE-oriented palaeovalleys, exhibiting depths at the seafloor of 4–7 m and widths of 1–3 km. These palaeovalleys appear to originate from about the position where the British ice-sheet southern front was at ~15 ka BP (Dove et al., 2017). The southern continuation of Pvs9 is obscured by the Norfolk Bank. Based on their orientation, it is however likely that these palaeo-rivers ran southward and into the NAic–Aic–Lic palaeo-river during ice melting.

HC is a 1–2 km wide, 5–7 m deep, NNW–SSE-oriented partially infilled palaeovalley (Figure 7.14). Radiocarbon dating of fluvial sediments infilling this palaeovalley yielded ages of 11,000–7,000 years BP (Limpenny et al., 2011). The southern end of HC is not clear in the bathymetry. Nevertheless, according to its orientation, this palaeovalley runs most likely into the southern channels of Pvs7. Consequently, the southern channels of Pvs7 may have been reactivated in early Holocene times by water coming from the HC. This may explain why the supposedly Late Saalian Pvs7 palaeovalley is not overhanging the possibly younger NAic palaeovalley.

7.5. Discussion

The geomorphological analysis of the southern North Sea seafloor has revealed several palaeovalleys that were previously unknown. This study has also helped to map more accurately the geomorphology of those previously identified, as well as to better establish their evolution, interrelationship and relative ages. The accurate mapping of these features leads to corroborate some hypotheses previously postulated to explain the Middle–Late Pleistocene fluvial/glacial evolution of this area. For instance, our analysis supports the glacial origin of the IGD and the southward deviation of the Medway–Thames palaeo-river during the Elsterian glaciation (Figure 7.15). Our study also provides evidences demonstrating the existence of a major palaeovalley system coming from northwestern Germany. In addition, the geophysical investigation have allowed to better constrain the last phases of evolution of the Axial–Rhine–Meuse–Thames palaeovalley system (Figure 7.17 and Figure 7.18).

Here below, we discuss the main geomorphological features described in this study and the palaeogeographic context in which they were likely formed.

7.5.1. Thames Estuary and Outer Thames Estuary palaeovalleys

The Outer Thames Estuary area exhibits a series of palaeovalleys and palaeo-depressions with different orientations. Several of them, such as Pvs2, Pv4, Pvs6 and Ed3, had never been identified before (Figure 7.6). The morphology of the seafloor in that area corroborates the presence of the east-trending palaeovalley (Pv3) previously associated with the pre-Elsterian Thames–Medway palaeo-river system (e.g. Bridgland and D'Olier, 1995). This palaeovalley is indeed one of the oldest features identified in the OTE. However, contrarily to previous geomorphological maps, our study shows that Pv3 does not extend to the present-day coastline (Figure 7.6). Rather, another palaeovalley (Pvs6) runs eastward from the Orwell–Stour River mouth, bending southward a few kilometers from the western termination of Pv3. Pvs6 has not been mapped prior to this study. Bridgland and D'Olier (1995) suggested though that a palaeovalley, which they called "the Clacton Channel", runs southeastward in that area. They interpreted that palaeovalley as a remainder of the Elsterian southward diverted Thames–Medway palaeo-river system. This interpretation may be valid for the initial incision of Pvs6, although some of the channels composing this system (e.g. v3) were most likely formed by more recent erosional events. Indeed, the close location of its western end to the mouth of present-day British river systems discharging into the southern North Sea suggests that Pvs6, or some of its palaeovalleys, may have been reused, or overrun, several times during subsequent marine lowstands.

Another enigmatic major erosional feature incised in this area are the IGD. This study provides the first detailed description of their morphology. The IGD appear to be part of a system (Ed1–IGD) comprising kilometer-scale, N-S-oriented elongated palaeo-depressions link to one another by palaeo-channels. This system seems to be independent from the rest of the palaeovalleys identified in the OTE. The particular configuration of the Ed3–IGD system, i.e. palaeo-depressions interconnected or ending in fluvial systems, resembles strongly the Weichselian elongated deeps incised in the Spur area, the formation of which is associated with glacial erosion (Figure 7.4; Dove et al., 2017). Taken together, their similarities to incisions typically found in palaeo-terrains eroded by glaciers, the fact that Ed1–IGD does not seem to have been carved within a palaeovalley, their isolation, and the homogeneity of the substratum into which they are carved, which excludes differential erosion, suggest that the system Ed3–IGD was formed by glacial erosion. The presence of an overdeepened depression at the base of the northern head of IGD and the palaeo-channels connecting Ed3 and IGD suggest the occurrence of one or several phases of fluvial erosion following their initial incision (Figure 7.9).

The possible glacial origin of the Ed3–IGD implies that the British Elsterian ice sheet may have extended several kilometers further south than previously thought (see also Emu Ltd & University of Southampton, 2009). The restriction of the deeps Ed3–IGD and the presence of Pvs6 in the west suggest that the possible ice sheet that extended along the OTE did not reach the present-day coastline. Rather, its eastern margin possibly extended along or nearby the eastern slope of Pvs6 (). The morphology of the eastern margin of that ice sheet is however unknown, as the various intense valley incisions that shaped the Axial Channel during subsequent marine lowstands may have erased any erosional feature formed by glacial erosion.

An ice sheet extending along the IGD might actually be the reason why the pre-Elsterian Medway–Thames palaeo-river (Pv3) was diverted southward during the Elsterian glacial Maximum (Figure 7.15). Indeed, a southward expansion of the glacier along offshore East Anglian and the OTE would have trapped the eastern half of Pv3 under the ice, blocking the eastern drainage route of that system. This would have forced its western part to flow southward along the ice margin, thus forming Pvs6. An ice sheet covering the OTE would also explain how Pvs2 and the deeps Ed2 formed. That is, seasonal melting of the ice-sheet front would have formed several fluvial systems at that location. That would indeed explain why Pvs2 has no apparent connection with any other palaeovalley identified in that area. The fact that Pvs2 seems to be truncated by the supposedly older Pv3 can be explained by a reusing of the eastern end of Pv3 by runoff once the ice sheet retreated from that area. Actually, some of the palaeo-channels linked to Ed3, which may represent glacial meltwater channels, seem to run into the easternmost part of Pv3 (Figure 7.6), indicating that some water was channeled through the eastern end of Pv3 when these palaeo-channel were active.

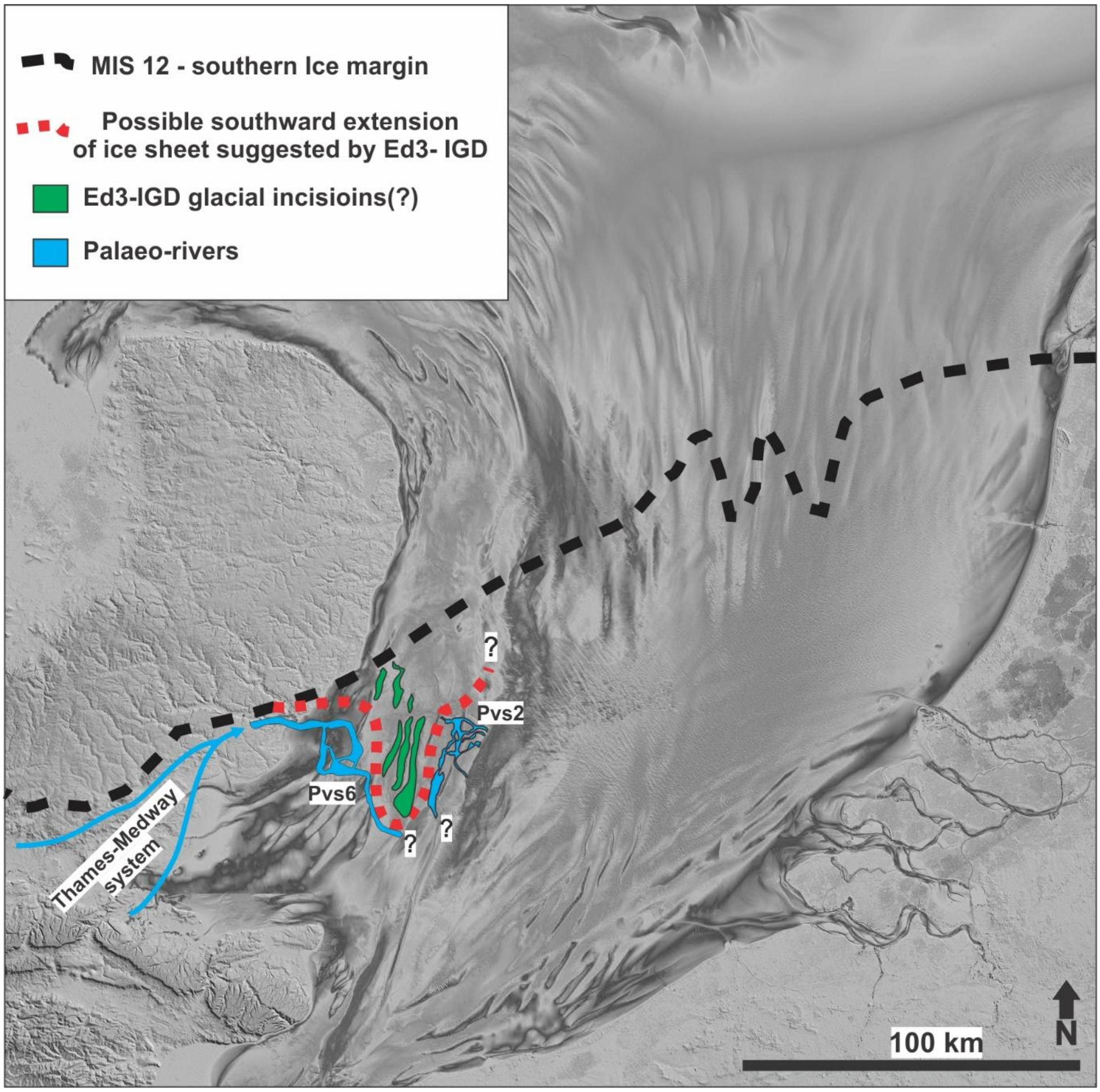

Figure 7.15. Possible palaeovalleys and deeps formed during the Elsterian glaciation identified in this study. Ice-sheet margin at Elsterian glacial maximum proposed by previous studies and hypothetic ice expansion along the OTE (this study) are indicated.

The datasets available for the present study do not permit to accurately map the entire buried Thames palaeovalley system extending along the TE. However, the geophysical data show that it comprises at least two main palaeovalleys carved and infilled at different times (Figure 7.11). In fact, the geophysical investigations suggest a southward migration of the Thames palaeo-river, as the younger palaeovalley (ThP2) is not carved within the older one (ThP1), but to the south of it. This corroborates the hypothesis put forward by Bridgland and D'Olier (1995), who suggested the presence of two palaeo-channels at the eastern end of the Thames palaeovalley system of Weichselian age. However, our investigation goes further, not only corroborating the presence of these two palaeovalleys, but suggesting that the incision and infilling of the northern palaeovalley (ThP1) took place before the incision of the southern one (ThP2). Based on previous interpretations suggesting that the Thames palaeo-river did not extend this far south until Late Saalian and/or Weichselian times (e.g. Gibbard, 1988; Gibbard, 1995; Bridgland and D'Olier, 1995) and the fact that ThP1 passed through a major phase of infilling before ThP2 was incised, we propose that ThP1 was formed during Late Saalian (Figure 7.16) and infilled during the Eemian highstand (130–110 ka BP). If that is so, ThP2 would represent the Thames palaeo-river during the Weichselian glaciation (Figure 7.17 and Figure 7.18). Based on this possible sequence of events and ThP2's cross-bedded seismic facies, we interpret its infill as fluvial sediments, and the internal erosional surfaces truncating it as changes in the river dynamics, such as variations in the volume of flow and/or in the river base level, produced at different stages of the Weichselian glaciation.

Palaeovalleys Pvs6 and Pvs8 were most likely reused or/and overprinted by younger river systems during the various Middle–Late Pleistocene marine lowstands that followed their formation. They are indeed prolongations of major on-land rivers (Figure 7.6). These palaeovalley systems or some of their channels might thus represent tributaries of the Thames palaeo-river during Saalian and Weichselian lowstands (Figure 7.17; see also Bridgland and D'Olier, 1995).

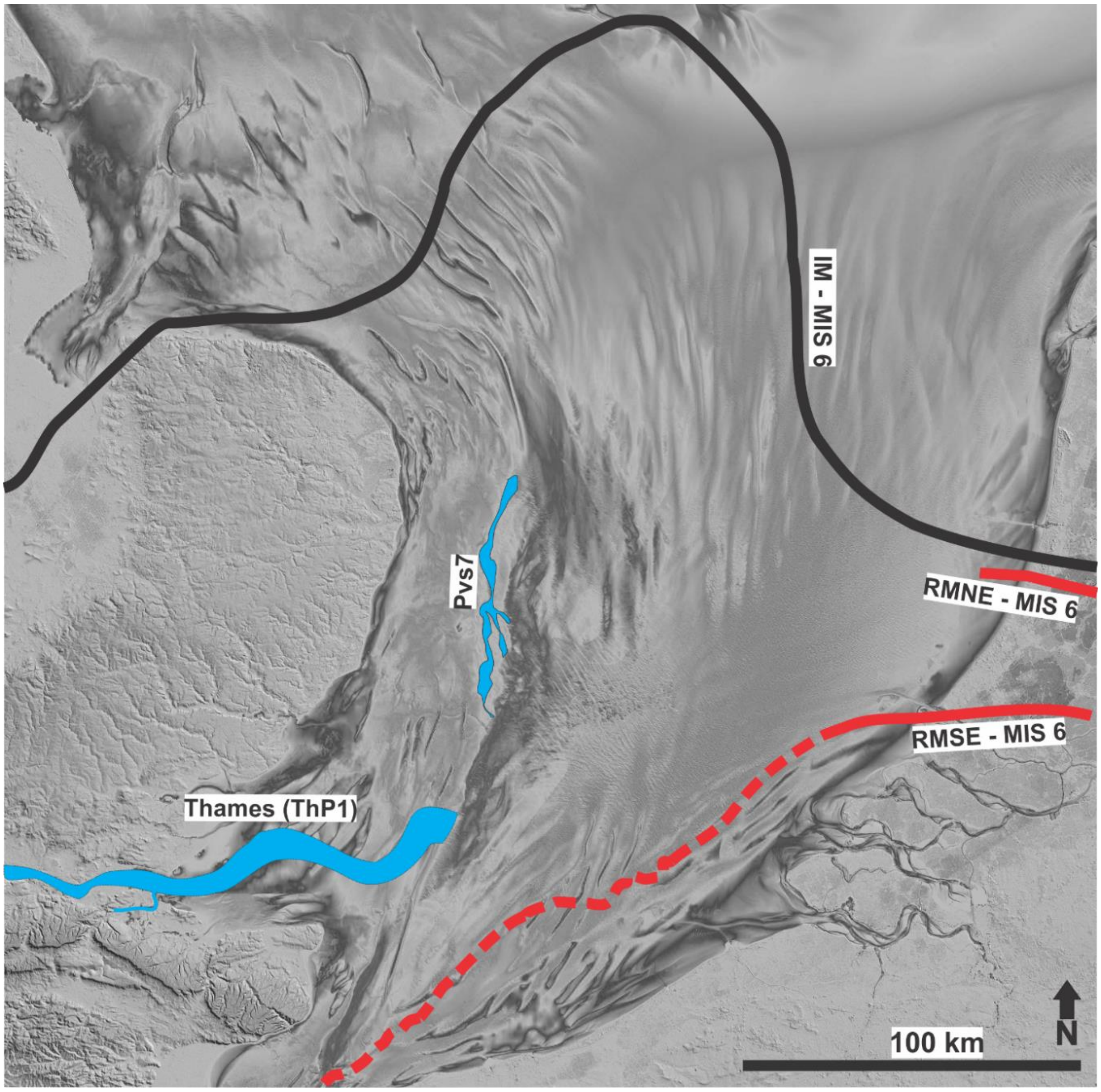

Figure 7.16. Possible palaeovalleys (blue areas) and scarps (dashed red line) carved during the Saalian glaciation interpreted from the data discussed in this study. Ice-sheet southern margin at Saalian glacial maximum (black line labelled IM – MIS 6) is indicated. RMNE–MIS 6: Saalian Rhine–Meuse northern edge; RMSE–MIS 6: Saalian Rhine–Meuse southern edge (see Figure 7.1).

7.5.2. Axial–Rhine–Meuse–Thames fluvial system

The drainage systems traversing the southern North Sea during the Saalian glaciation are poorly preserved in the seafloor. Only palaeovalleys Pvs7 and ThP1 appear to have been carved during that period, corroborating that palaeo-rivers did traverse the southern North Sea at Saalian glacial maximum. Previous studies have also associated the formation of scarp S4 with Late Saalian fluvial erosion. In particular, this scarp is interpreted as the southern palaeo-slope of the Rhine–Meuse palaeo-river system, which purportedly extended over this area during the Saalian glacial maximum (e.g. Mathys, 2009). S4 appears to align with (and to have similar height of) scarp E2 in the Dover Strait, suggesting that both scarps may be parts of the same palaeo-slope (Figure 7.16). The possible continuation of S4 across the Dover Strait thus supports the model in which the Rhine–Meuse palaeo-river system ran southwestward during the Saalian glaciation, entering the English Channel through the Dover Strait. For this to happen, the Dover Strait should have been, at least, partially opened by then, which is consistent with an Elsterian breach of the Weald–Artois ridge. Note that none of the possible Saalian palaeovalleys and other fluvial erosional features identified in the southern North Sea are incompatible with the presence of a proglacial lake during the Saalian glacial maximum.

The Weichselian palaeo-river systems appear to be better preserved in the southern North Sea. The geomorphology of the seafloor and the top pre-Quaternary surface underneath the Holocene cover in the BCS indicates that the Rhine–Meuse palaeo-river system was channeled through the Axial Channel during the Weichselian glaciation. However, the morphology of these palaeovalleys reveals that the Axial Channel was not only the westward prolongation of the Rhine–Meuse palaeo-river system, but a major river formed by the convergence of the latter with the Thames and a major fluvial system coming from the north (i.e. the North Axial Channel or NAC; see Figure 7.17).

Based on the extent and orientation of the NAC, we associate its formation with the major drainage system that, according to Sejrup et al. (2016) and Toucanne et al. (2015), among others, flowed southward from the German–Danish continental shelf 30–18 Ka BP. The geophysical data available for this study does not give any information on whether that palaeo-river was the outflow of an ice-marginal lake or the result of converging rivers. The bathymetric data suggests though that the location proposed by Sejrup et al. (2016) for the formation of a proglacial lake appears to have lower relief than its surroundings. In addition, the bathymetry of that area does not exhibit any geomorphological feature suggesting that major palaeovalleys cut through it. The proglacial lake hypothesis thus seems to correspond better with the geomorphology of the seafloor in that area. However, the lack of seismic reflection data from that location prevents the corroboration of that hypothesis.

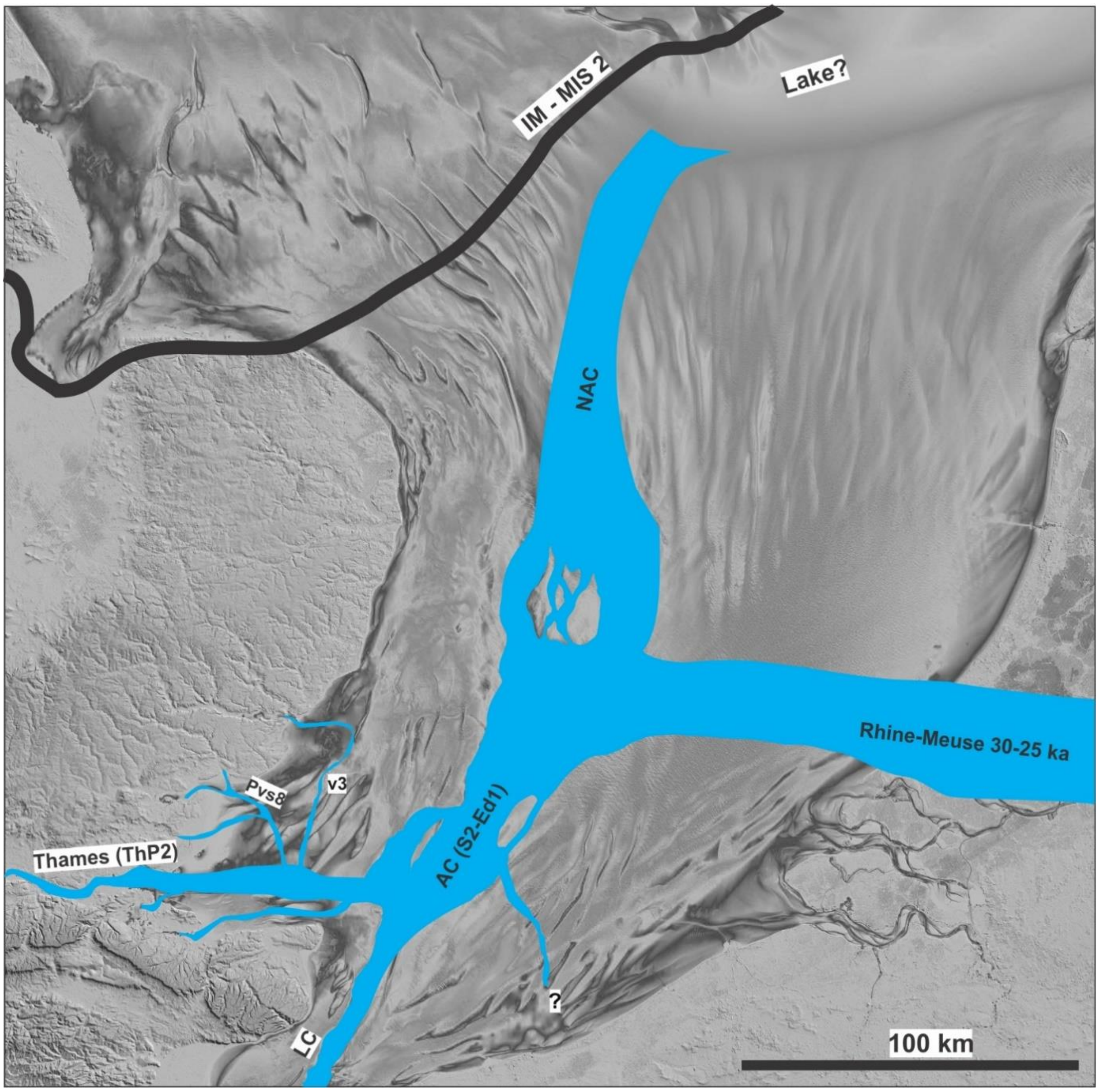

Figure 7.17. Possible palaeovalleys (blue areas) incised during the Weichselian glacial maximum interpreted from the data discussed in this study. Ice-sheet margin at Weichselian glacial maximum (black line labelled IM – MIS 2) is indicated (see Figure 7.1). Labels are explained in previous figures and main text.

The morphology of the NAC is poorly constrained. The geomorphology of the seafloor suggests though that its width might have covered the area from the western edge of the NAic to the position of Ed5; that is, a width of up to ~45 km (Figure 7.17). However, more data are needed to demonstrate that, as the origin of Ed5 is unknown. Indeed, the possible glacial origin of the IGD casts doubt on the currently accepted position of the maximum southern expansion of ice sheets during the Elsterian Glaciation. We therefore cannot exclude that Ed5 were formed due to glacial erosion. Actually, at this stage, we cannot rule out Saalian fluvial erosion. Neither can we exclude Eemian and/or Holocene submarine sedimentary/erosional processes.

The geomorphology of the area where the Rhine–Meuse, the NAC and the Axial Channel palaeovalleys converge is also unclear in the bathymetry due to the presence of extensive dune fields. The orientations of scarps S2 and S3 appear to be consistent with the position of the southern slope of the Rhine–Meuse palaeo-river system proposed for 30–25 ka BP (e.g. Liu et al., 1992; Hijma et al., 2012). Based on that, we propose that the valley incisions along the Axial Channels defined by S2–Ed1 and S3–Ed1 correspond to two erosional episodes that occurred during the 30–25 ka phase of the Rhine–Meuse palaeo-river system (Figure 7.17). The path that the NAC–Rhine–Meuse–Thames–Axial Channel palaeovalley system followed across the Dover Strait is not distinct in the bathymetry. Indeed, scarp S3 is truncated by S2 tens of kilometers northward of the Dover Strait, and S2 appears to converge with the Lic (Figure 7.13), which most likely truncates it. This suggests that none of the scarps identified in the Dover Strait (i.e. E1 and E2) correspond to any of those phases of valley incision.

The Lic–Aic–NAic palaeovalley represents the youngest fluvial erosion observed in the NAC–Axial–Lobourg Channel (Figure 7.18). That is evidenced by the fact that Pvs1, which runs into the Lic, truncates all fluvial geomorphological features formed before the Lic in the Dover Strait. This demonstrates that, at the time that Pvs1 hosted a river, the flow along the Axial–Lobourg Channel was channeled through the Lic–Aic–NAic. That is also consistent with some of the apparent palaeo-channels identified along the possible western continuation of the Rhine–Meuse palaeovalley, which appear to terminate at their intersection with Aic–NAic (Figure 7.13). The Lic also seems to truncate Ed1 and S2 indicating that this valley was eroded due to a later event.

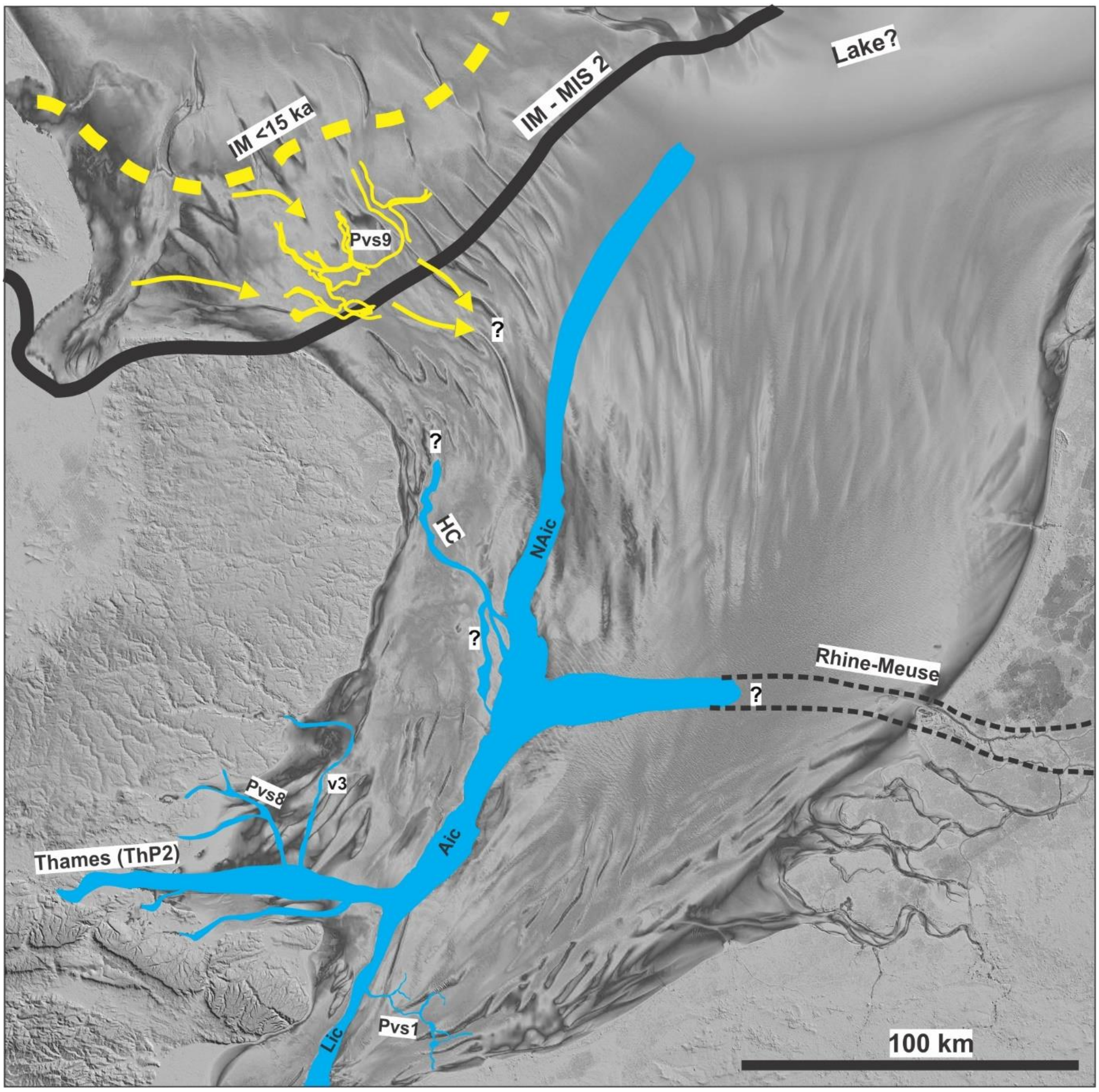

Figure 7.18. Palaeovalleys (blue areas) representing the last phase of valley incision identified in the study area according to interpretations of the data discussed in this study. Ice-sheet southern margins at Weichselian glacial maximum (black line labelled IM – MIS 2) and at ~15 ka BP (yellow line labelled IM> 15 ka; Dove et al., 2017) are indicated. Labels are explained in previous figures and main text. We have also included palaeovalley HC, as, even though its infill has yield early Holocene ages, it is quite likely that it hosted a river during the Weichselian glaciation too.

The erosional processes that carved the Lic–Aic–NAic inner channel seem to have been extremely intense. This is evidenced by the fact that this palaeo-channel runs across a range of different substratum units (sands, clays, chalk, etc.), keeping a linear course and a general box-shape cross-sectional morphology regardless the nature of the substratum. This geomorphology is indeed similar to those found in high energy fluvial systems and/or along megaflood-eroded valleys (e.g. Kehew and Lord, 1986; Wohl, 1993; Rains et al., 1993; Kale et al., 1996; Baynes et al., 2015). In addition, the Lic comprises a range of bed-form erosions typically found in megaflood-eroded terrains (Chapters 5 and 6; see also Rains et al., 1993; Lamb et al., 2007; Baker, 2009; Shaw, 2010; Baynes et al., 2015). Hence, it is very unlikely that the concentration of the erosion along that channel happened because progressive river erosion only. Alternatively, we propose that the NAic–Aic–Lic was incised by one or several megaflood event(s) and modified by subsequent progressive fluvial erosion. The source of these events is however unknown. They may have been generated either by lake-outburst flood flows produced at the proglacial lake hypothesized by Sejrup et al. (2016), or by lake outburst floods or massive inputs of glacial meltwater discharged into the rivers that run into the German–Danish continental shelf of the North Sea during the Last Glacial Maximum (Toucanne et al., 2010; Toucanne et al., 2015; Patton et al., 2017). The data available for this study are insufficient to determine when this flooding episode(s) occurred. However, the NAic–Aic–Lic inner channel appears to be younger than the palaeovalley purportedly representing the 30–25 ka Rhine–Meuse palaeo-river system. That suggests that the megaflood that incised the inner channel occurred during later stages of the Weichselian glaciation. It is thus possible that the incision of this palaeovalley is related to one or several of the significant meltwater discharges from the Channel palaeo-river that took place 22.7–21.1 ka, 20.5–18.4 ka and/or 18.4–16.5 ka BP (Toucanne et al., 2015).

7.6. Conclusions

Thanks to the extensive geophysical dataset gathered for this study, we have been able to better characterize the geomorphology and relative ages of the various palaeovalleys imprinted on the seafloor of the southern North Sea. We have also identified several palaeovalleys never mapped before. In addition, the detailed geomorphological interpretation of the different palaeovalleys and palaeo-depressions carved in the southern North Sea has led to support a series of hypotheses previously postulated with geomorphologic evidence.

Firstly, we demonstrate the existence of a major palaeo-drainage system that ran southward from the German–Danish continental shelf during the Last Glacial Maximum; i.e. the North Axial Channel. The North Axial Channel converges with the Rhine–Meuse and Thames palaeovalleys, entering the English Channel through the Dover Strait as a single palaeovalley; i.e. the Axial–Lobourg Channel. This palaeovalley system was thus the northern continuation of the Channel palaeo-river system, at least during the Last Glacial Maximum.

We have found evidence suggesting the occurrence of 2–3 phases of valley incision along the Axial–Lobourg Channel during the Weichselian glacial maximum. Based on our geomorphological analysis and palaeogeographic reconstructions, we postulate that the oldest ones took place 30–25 ka BP. The youngest one (Lic–Aic–NAic), on the other hand, seems to be linked to later erosional events, although still within the timeframe during which German palaeo-rivers were southward diverted (i.e. 30–18 ka BP). More importantly, the youngest valley incision resulted in the formation of the inner channel NAic–Aic–Lic, which exhibits characteristics similar to megaflood-eroded valleys. This channel thus indicates the occurrence of high-magnitude flood flows in the southern North Sea and eastern English Channel during the Weichselian glacial maximum.

Secondly, the geomorphological characterization of the Outer Thames Estuary and East Anglia areas supports glacial erosion as the main erosional agent that carved palaeo-depressions Ed3 and IGD. This means that the British–Irish Elsterian ice sheet may have extended along East Anglia and the OTE several kilometers further to the south than previously thought. Our study also shows that the pre-Elsterian Thames–Medway palaeovalley imprinted on the bathymetry does not extend to the British coastline. Rather, the palaeovalley extending eastward from the coastline bends southward several kilometers before attaining the latter. This palaeovalley may represent the valley carved by the southward diversion of the Thames–Medway palaeo-river during the Elsterian glacial maximum. More importantly, the southward deviation of that palaeo-river system may be a direct consequence of the possible southward expansion of the ice sheet along the OTE during that glacial age.

Last but not least, our study shows the presence of two main buried palaeovalleys at the eastern end of the Thames palaeovalley system, which extend eastward across the Thames Estuary from its present-day mouth. We interpret these palaeovalleys as the eastern continuation of the Thames paleo-river during Late Saalian and Weichselian glacial periods.

In conclusion, this study significantly contributes to the understanding of the evolution of the main European fluvial systems flowing through the southern North Sea during major Pleistocene glaciations. It also provides important input on the extents of the ice sheets during the Elsterian glacial maximum, which, according to our findings, should be revised. However, many unknowns still remain on the palaeovalleys' geomorphologies and ages. Acquisition of additional geophysical and geological data, in combination with data existing in other European and commercial databases, is key to better reconstruct the evolution of these palaeovalleys.

References

Baker, V.R., 2009, The Channeled Scabland: A Retrospective, Annual Review of Earth and Planetary Sciences, v. 37, p. 393–411.

Balson, P.S., and D'Olier, B.D., 1989, Thames estuary Sheet 51°N–00°, solid geology, British Geological Survey, 1:250 000 series, 1 sheet.

Balson, P.S., and D'Olier, B.D., 1990, Thames estuary sheet 51°N–00°, seabed sediments and Quaternary geology, British Geological Survey, 1:250 000 series, 1 sheet.

Balson PS, Laban C, Frantsen PJ, Parker, N., Henriet, J.P., De Batist and M., Mostaert, F, 1991. Ostend Sheet 51°N–02°E, Solid geology, British Geological Survey, 1:250.000 series, 1 sheet.

Baynes, E. R. C., Attal, M., Dugmore, A. J., Kirstein, L. A. and Whaler, K. A., 2015, Catastrophic impact of extreme flood events on the morphology and evolution of the lower Jokulsa a Fjollum (northeast Iceland) during the Holocene, Geomorphology, v. 250, p. 422–436.

Behre, K-E., 2007, A new Holocene sea-level curve for the southern North Sea, Boreas, v. 36, p. 82–102.

Bateman, M.D., Buckland, P.C., Chase, B., Frederick, C.D., and Gaunt, G.D., 2008. The late-Devensian proglacial Lake Humber: new evidence from littoral deposits at ferrybridge, Yorkshire, England, Boreas, v. 37, p. 195–210.

Bateman, M.D., Buckland, P.C., Whyte, M.A., Ashurst, R.A., Boulter, C., and Panagiotakopulu, E.V.A., 2011, Re-evaluation of the last glacial maximum typesite at Dimlington, UK, Boreas, v. 40, p.573–584.

Bridgland, D.R., and D'Olier, B., 1995, The Pleistocene evolution of the Thames and Rhine drainage systems in the southern North Sea Basin, in: Preece, R. C. (ed.), 1995, Island Britain: a Quaternary perspective Geological Society Special Publication, v. 96, p. 27-45

Busschers, F. S., Weerts, H. J. T., Wallinga, J., Cleveringa, P., Kasse, C., De Wolf, H., and Cohen, K. M., 2005, Sedimentary architecture and optical dating of Middle and Late Pleistocene Rhine-Meuse deposits-fluvial response to climate change, sea-level fluctuation and glaciation, Netherlands Journal of Geosciences, v. 84, p. 25–41.

Busschers, F. S., Kasse, C., Van Balen, R. T., Vandenberghe, J., Cohen, K. M., Weerts, H. J. T., Wallinga, J., Johns, C., Cleveringa, P., and Bunnik, F. P. M., 2007, Late Pleistocene evolution of the Rhine-Meuse system in the southern North Sea basin: imprints of climate change, sea-level oscillation and glacio-isostacy, Quaternary Science Reviews, v. 26, p. 3216–3248

Cameron, T.D.J.; Laban, C., Schüttenhelm, R.T.E., 1984a, Flemish Bight sheet 52N–02E, Quaternary Geology, 1:250 000 Series, British Geological Survey, 1 sheet.

Cameron, T.D.J.; Grimshaw, S., Hall, D.M., Laban, C., Mesdag, C.S., Parker, N., and Schüttenhelm, R.T.E., 1984b, Flemish Bight sheet 52N–02E, Solid Geology, 1:250 000 Series, British Geological Survey, 1 sheet.

Cameron, T.D.J., Laban, C., Mesdag, C.S., and Schüttenhelm, R.T.E., 1986, Indefatigable sheet 53N–02E, Quaternary Geology, 1:250 000 Series, British Geological Survey, 1 sheet.

Cameron, T. D. J., Stoker, M. S., and Long, D., 1987a, The history of Quaternary sedimentation in the UK sector of the North Sea Basin, Journal of the Geological Society, v. 144, p. 43–58.

Cameron, T.D.J., Bulat, J., Parker, N., Frantsen, P., Laban, C., and Schüttenhelm, R.T.E., 1987b, Indefatigable sheet 53N–02E, Solid Geology, 1:250 000 Series, British Geological Survey, 1 sheet.

Carr, S.J., Holmes, R., van der Meer, J.J.M., Rose, J., 2006, The last glacial maximum in the North Sea basin; micromorphological evidence of extensive glaciation, Journal of Quaternary Science, v. 21, p. 131–153.

Caston, V. N. D., and Stride, A. H., 1973, Influence of older relief on the location of sand waves in a part of the southern North Sea, Estuarine and Coastal Marine Science, v.1, p. 379–238.

Clark, C.D., Gibbard, P.L. and Rose, J., 2004, Pleistocene glacial limits in England, Scotland and Wales, in Ehlers, J., and Gibbard, P.L., Eds., Developments in Quaternary Sciences. Elsevier, p. 47–82.

Cohen, K.M., Gibbard, P.L., and Weerts, H.J.T., 2014, North Sea palaeogeographical reconstructions for the last 1 Ma, Netherlands Journal of Geosciences – Geologie en Mijnbouw, v. 93, p. 1–23.

Collier, J. S., Oggioni, F., Gupta, S., García-Moreno, D., Trentesaux, A., De Batist, M., 2015, Streamlined islands and the English Channel megaflood hypothesis, Global and Planetary Change, v. 135, p. 190–206.

De Clercq, M., ChaDTMenos, V., Van Lancker, V., & Missiaen, T., 2016, A high-resolution DTM for the Top-Palaeogene surface of the Belgian Continental Shelf, Journal of Maps, v. 12, p. 1047–1054.

Dove, D., Evans, D. J.A., Lee, J.R., Roberts, D.H., Tappin, D. R., Mellett, C.L., Long, D., and Callard, S.L., 2017, Phased occupation and retreat of the last British–Irish Ice Sheet in the southern North Sea; geomorphic and seismostratigraphic evidence of a dynamic ice lobe, Quaternary Science Reviews, v. 163, p. 114–134.

Dyer, K.R., and Huntley, D.A., 1999, The origin, classification and modelling of sand banks and ridges, Continental Shelf Research, v.19, p. 1285–1330.

Ehlers, J., and Gibbard, P. L. (Eds.), 2004, Quaternary glaciations-extent and chronology: part I: Europe. Developments in Quaternary Science, vol. 2, Elsevier, Amsterdam.

Emu Ltd and University of Southampton, 2009, The Outer Thames Estuary Regional Environmental Characterisation, Final Report, number 09/J/1/06/1305/0870, Published by

Marine Aggregate Levy Sustainability Fund. Available from http://www.marinedataexchange.co.uk

Gaunt, G.D., Crosby, A., and Hall, D.M., 1985, Spurn solid geology, Sheet 53°N–00°, British Geological Survey, scale 1:250 000, 1 sheet.

Gibbard, P.L., 1995, The formation of the Strait of Dover, Geological Society, London, Special Publications, v. 96, 15–26.

Gibbard, P.L., and Cohen, K.M., 2015, Quaternary evolution of the North Sea and the English Channel, Proceedings of the Open University Geological Society, v. 1, p. 63–74

Gupta, S., Collier, J.S., García-Moreno, D., Oggioni, F., Trentesaux, A., Vanneste, K., De Batist, M., Camelbeeck, T., Potter, G., Van Vliet Lanoe, B., and Arthur, J.C.R., 2017, Making Britain: Two-stage catastrophic opening of the Dover Strait, Nature Communications, DOI: 10.1038/ncomms15101.

Hamblin, R.J.O., Crosby, A., Balson, P.S., Jones, S.M., Chadwick, R.A., Penn, I.E., and Athur, M.J., 1992, United Kingdom offshore regional report: the geology of the English Channel, London: HMSO for the British Geological Survey.

Hijma, M. P., Cohen, K. M., Roebroeks, W., Westerhoff, W. E., and Busschers, F. S., 2012, Pleistocene Rhine–Thames landscapes: geological background for hominin occupation of the southern North Sea region, Journal of Quaternary Science, v. 27, p. 17–39.

Henriet, J.P., De Moor, G., and De Batist, M. (eds.), 1989, The Quaternary and Tertiary Geology of the Southern Bight, North Sea, International Colloquy on the Quaternary and Tertiary Geology of the Southern Bight, North Sea, Ghent, Belgiumn May 28–30, 1984, 241 pages.

Hopson, P.M., Zalasiewicz, J.A., and Harrison, D.J., 1991, East Anglia sheet 52°N–00°, Quaternary Geology, British Geological Survey, 1:250 000 series, 1 sheet.

Houbolt, J.J.H.C., 1968, Recent sediments in the southern bight of the North Sea, Geologie en mijnbouw, v. 47, p. 245–273.

James, J.W.C., Guennoc, P., Harrison, M., Le Bot, S., Philpott, S., Vinchon, C., Bee, E., Simien, F., Janjou, D., Garlan, T., Trentesaux, A., Mahieux, G., Briet D. and, Augris C., 2002, GEOSYNTH: A synthesis of the geology and sediments of the Dover Strait and its hinterland. CR/02/078, CDROM, British Geological Survey.

Kehew, A.E. and Lord, M.L., 1986, Origin and large-scale erosional features of Glacial-Lake spillways in the northern Great-Plains. Geological Society of America Bulletin, v. 97, p. 162–177.

Laban, C., Schüttenhelm, R.T.E., Balson, P.S., Baeteman, C., Paepe, R., 1992. Ostend Sheet 51N–02E. 1:250 000 Series. Quaternary Geology. British Geological Survey – Rijks Geologische Dienst – Belgische Geologische Dienst. Nottingham – Haarlem – Brussel.

Lamb, M.P., Howard, A.D., Dietrich, W.E. and Perron, J.T., 2007, Formation of amphitheater-headed valleys by waterfall erosion after large-scale slumping on Hawaii. Geological Society of America Bulletin, v. 119, p.805–822.

Le Bot, S., Van Lancker, V., Deleu, S., De Batist, M., and J.P. Henriet, 2003, Tertiary and Quaternary Geology of the Belgian Continental Shelf, SPSD II report, 77 pages.

Lee, J. R., Busschers, F. S., and Sejrup, H. P., 2012, Pre-Weichselian Quaternary glaciations of the British Isles, The Netherlands, Norway and adjacent marine areas south of 68 N: implications for long-term ice sheet development in northern Europe, Quaternary Science Reviews, v. 44, p. 213–228.

Lewis S., Maddy D., and Glenday S., 2004, The Thames valley sediment conveyor: fluvial system development over the last two interglacial-glacial cycles. Quaternaire, v. 15, p. 17–28.

Limpenny, S.E., Barrio Froján, C., Cotterill, C., Foster-Smith, R.L., Pearce, B., Tizzard, L., Limpenny, D.L., Long, D., Walmsley, S., Kirby, S., Baker, K., Meadows, W.J., Rees, J., Hill, J., Wilson, C., Leivers, M., Churchley, S., Russell, J., Birchenough, A.C., Green, S.L., and Law, R.J., 2011, The East Coast Regional Environmental Characterisation, Cefas Open report 08/04 (available at www.alsf-mepf.org.uk), 287 pages.

Liu, A. C., Missiaen, T., and Henriet, J. P., 1992, The morphology of the top-Tertiary erosion surface in the Belgian sector of the North Sea, Marine Geology, v.105, p. 275–284.

Liu, A. C., De Batist, M., Henriet, J. P., and Missiaen, T., 1993, Plio-Pleistocene scour hollows in the Southern Bight of the North Sea, Geologie en Mijnbouw, v. 71, p. 195–204.

Mathys, M., 2009, The Quaternary geological evolution of the Belgian Continental Shelf, southern North Sea, PhD thesis, Ghent University, Belgium.

Mellett, C., Hodgson, M., Plater, A., Mauz, B., Selby, I. & Lang, A., 2013, Denudation of the continental shelf between Britain and France at the glacial–interglacial timescale, Geomorphology, v. 203, p. 79–96.

Patton, H., Hubbard, A., Andreassen, K., Auriac, A., Whitehouse, P. L., Stroeven, A. P., Shackleton.C., Winsborrow, M., Heyman, J., and Hall, A. M., 2017, Deglaciation of the Eurasian ice sheet complex, Quaternary Science Reviews, v. 169, p. 148–172.

Peeters, J., Busschers, F. S., and Stouthamer, E., 2015, Fluvial evolution of the Rhine during the last interglacial-glacial cycle in the southern North Sea basin: a review and look forward. Quaternary International, v. 357, p. 176–188.

Peeters, J., Busschers, F. S., Stouthamer, E., Bosch, J. H. A., Van den Berg, M. W., Wallinga, J., Versendaal, A.J., Bunnik, F.P.M., and Middelkoop, H., 2016, Sedimentary architecture and chronostratigraphy of a late Quaternary incised-valley fill: A case study of the late Middle and Late Pleistocene Rhine system in the Netherlands, Quaternary Science Reviews, v. 131, p. 211–236.

Rains, B., Shaw, J., Skoye, R., Sjogren, D., and Kvill, D., 1993, Late Wisconsin subglacial megaflood paths in Alberta. Geology, v. 21, p. 323–326.

Sejrup, H. P., Nygård, A., Hall, A. M., and Haflidason, H., 2009, Middle and Late Weichselian (Devensian) glaciation history of south-western Norway, North Sea and eastern UK, Quaternary Science Reviews, v. 28, p. 370-380.

Sejrup, H. P., Clark, C. D., and Hjelstuen, B. O., 2016, Rapid ice sheet retreat triggered by ice stream debuttressing: Evidence from the North Sea. Geology, v. 44, p. 355–358.

Shaw, J., 2010, In defence of the meltwater (megaflood) hypothesis for the formation of subglacial bedform fields, Journal of Quaternary Science, v. 25, p. 249–260.

Tappin, D.R., 1991, Spurn Quaternary geology, Sheet 53°N–00°, British Geological Survey, scale 1:250 000, 1 sheet.

Toucanne, S., Zaragosi, S., Bourillet, J. F., Gibbard, P. L., Eynaud, F., Giraudeau, J., Turon, J.L., Cremer, M., Cortijo, E., Martinez, P., and Rossignol, L., 2009a, A 1.2 Ma record of glaciation and fluvial discharge from the West European Atlantic margin. Quaternary Science Reviews, v. 28, p. 2974–2981.

Toucanne, S., Zaragosi, S., Bourillet, J.F., Cremer, M., Eynaud, F., Van Vliet-Lanoë, B., Penaud, A., Fontanier, C., Turon, J.L., Cortijo, E. and Gibbard, P.L., 2009b, Timing of massive 'Fleuve Manche' discharges over the last 350 kyr: insights into the European ice-sheet oscillations and the European drainage network from MIS 10 to 2, Quaternary Science Reviews, v. 28, p. 1238–1256.

Toucanne, S., Zaragosi, S., Bourillet, J. F., Marieu, V., Cremer, M., Kageyama, M., Van Vliet-Lanoë, B., Eynaud, F., Turon, J-L, and Gibbard, P. L., 2010, The first estimation of Fleuve Manche palaeoriver discharge during the last deglaciation: evidence for Fennoscandian ice sheet meltwater flow in the English Channel ca 20–18ka ago, Earth and Planetary Science Letters, v. 290, v. 459–473.

Toucanne, S., Soulet, G., Freslon, N., Jacinto, R. S., Dennielou, B., Zaragosi, S., Eynaud, F., Bourillet, J-F., and Bayon, G., 2015. Millennial-scale fluctuations of the European Ice Sheet at the end of the last glacial, and their potential impact on global climate. Quaternary Science Reviews, 123, 113-133.

Wohl, E.E., 1993, "Bedrock Channel Incision along Piccaninny Creek, Australia," The Journal of Geology, v. 101, p. 749–761.

Zalasiewicz, J.A., and Balson, P.S., 1985, East Anglia sheet 52°N–00°, Solid Geology, British Geological Survey, 1:250 000 series, 1 sheet.

Chapter 8

Discussion: Palaeogeographic reconstructions

Chapter 8 – Discussion: Palaeogeographic reconstructions

The geomorphological studies performed in the framework of this dissertation on the various exposed and buried palaeovalleys and palaeo-depressions incised in the southern North Sea and Dover Strait have permitted to better assess the processes and palaeogeographic settings in which these features were formed. Many of these features are tightly linked to (or attest to) the most significant palaeogeographic modifications that affected northwestern Europe during Middle–Late Pleistocene glacial periods. Hence, despite the absence of absolute dating and the many uncertainties still remaining, we can tentatively contextualize the formation of these erosional features within some of the palaeogeographic phases that occurred during each of the last three major Pleistocene marine lowstands.

The initial incisions of major palaeovalleys and palaeo-depressions carved in this area appear to be associated with intense fluvial and flooding erosional events that occurred during the various Middle–Late Pleistocene glacial maximum. Erosion during the glacial stages other than the glacial maximum and marine erosion during highstands were most likely significant too, since the southern North Sea was exposed to subaerial and subaquatic erosional settings over thousands of years. However, it was the southwestward diversion of southern North Sea palaeo-drainage systems during Middle–Late Pleistocene glacial maxima that caused the formation of major bedrock-incised palaeovalleys and palaeo-depressions. Nevertheless, progressive fluvial erosion during lowstands and marine erosion during highstands most likely contributed to the final shape of the palaeovalleys and palaeo-depressions carved during the glacial maxima. In addition, tidal erosional-and-depositional processes during marine transgressions and subsequent submarine and coastal erosion during highstands must have contributed significantly to the formation of the present-day Dover Strait. Of course, erosion and deposition in these areas during glacial and interglacial periods must have been significantly influenced by isostatic depression and rebound too. In Table 8.1, we summarize the various major palaeogeographic modifications that most likely caused the main erosional features identified in this study.

Event	Associated erosional features	Glacial/interglacial
Ice extending over offshore East Anglia and OTE	Incision of IGD and Thames southward deviation.	Elsterian Glaciation (478–424 ka BP)
Ice merging across the central North Sea and Inundation of the southern North Sea by a lake	Fosses Dangeard	Elsterian glacial max. (~450 ka BP)
Breach of the Weald-Artois ridge and first megaflood. Establishment of a river (outflow of the lake) in the center of the Strait.	Intense erosion of the Fosses Dangeard and possible incision of some of the Channel palaeovalleys by the megaflood and subsequent fluvial erosion.	Elsterian glacial max. (~450 ka BP).
Marine transgression and subsequent Highstand	Tidal and coastal erosion of the opening made in the Weald-Artois ridge, tidal sand ridge formation, infilling of the Fosses Dangeard.	Holsteinian interglacial (424–374 ka BP)
Ice merging across the central North Sea and inundation of part of the southern North Sea by a lake	Thames palaeo-river flowed through ThP1.	Saalian glacial max. (175–155 ka BP)
Erosion of the land-bridge damming the lake in the southern North Sea, inducing massive fluvial discharges in the Channel River (megaflood?). Establishment of a continuous river system (Thames-Rhine-Meuse-Channel palaeo-river) across the southern North Sea and English Channel	Incision of internal erosional surfaces Eb1 and/or Eb4 in the Fosses Dangeard's infill, reshaping and rerouting of palaeo-rivers composing the Channel River. Incision of escarpments S4 and palaeovalley Pvs7 in the southern North Sea, and possibly palaeovalleys LC1 and/or LC2 in the Dover Strait.	Saalian glacial max. (150–140 ka BP)
Marine transgression and subsequent Highstand	Possibly tidal, submarine and coastal erosion of the remnants of the Weald-Artois ridge and northern land-bridge, formation of sandbanks and dunes, last phases of infilling of Fosses Dangeard.	Eemian interglacial (130–110 ka BP)
Ice merging across the central North Sea and establishment of a continuous river system (Thames–Rhine-Meuse–North Axial–Axial–Channel palaeo-river) and a possible ice-marginal lake in the German–Danish Continental Shelf	Incision of the North Axial Channel, formation of inner channel (Naic-Aic-Lic) along the Lobourg-Axial Channel by a megaflood, possible incision or reshaping of the Northern Palaeovalley in the Channel. Thames palaeo-river flowed through ThP2.	Weichselian glacial maximum and subsequent episodes of massive glacial melting (30–16.5 ka BP)
Marine transgression and subsequent Highstand	Transgressional and submarine erosion of the seafloor. Coastal erosion of the Dover Strait, formation of sandbanks and dunes in the southern North Sea and English Channel.	Holocene interglacial (12 ka BP – present)

Table 8.1. Palaeogeographic events that most likely caused the formation of the major erosional features identified in the Dover Strait and southern North Sea.

Here below, we discuss the implications our findings have on Middle–Late Pleistocene palaeogeographic/palaeo-hydrographic reconstructions of the southern North Sea and eastern English Channel. Note that the models proposed below do not explain the entire glacial periods. Neither do we discuss palaeogeographic modifications during subsequent marine highstands, as, apart from the Holocene one, we have not been able to distinguish erosional/depositional features associated to any of these in the study area. We only discuss the palaeogeography during glacial maxima of the last three major Pleistocene glaciations, which are most likely the time intervals when the various erosional features identified in this study were formed.

8.1 Elsterian glacial maximum (~450 ka BP)

The geomorphology and spatial distribution of the various palaeo-depressions composing the Fosses Dangeard indicate that these features were carved by plunge-pool erosion at the base of northward retreating waterfalls. Our study thus corroborates the existence of a proglacial lake dammed at today's Dover Strait. Incidentally, this demonstrates that southeastern England and northern France were linked by a barrier that extended across the Dover Strait.

The morphology of the Fosses Dangeard and the erosional surfaces interrupting their infills suggest the occurrence of only one phase of plunge-pool erosion in the Dover Strait. This implies that the waterfall phase and the subsequent breach of the Weald–Artois ridge that initiated the opening of the Dover Strait should have happened during the same marine lowstand. On the other hand, the various scouring-and-infilling episodes interrupting the infills of the central palaeo-depressions may have occurred during subsequent lowstands and/or highstands. The absolute age of the formation and infilling of the Fosses Dangeard, and hence the timing of the opening of the Dover Strait, remain poorly constrained. Present consensus holds that the Dover Strait initially opened around 450 ka BP (e.g. Gibbard and Cohen, 2015).

The dimensions of the Elsterian southern North Sea proglacial lake remain unknown. Based on the height of the cliffs at Dover and Calais, and glacio-lacustrine deposits identified along East Anglia and in the North Sea (e.g. Gibbard, 1995), several authors (e.g. Murton and Murton, 2012; Gibbard and Cohen, 2015; Gibbard and Lewin, 2016) have theorized that the

Weald–Artois ridge may have reached heights of 30 m above present-day sea level. This, added to the depth of the present-day seafloor at the Dover Strait and the possible relative isostatic uplift of that area during this glaciation, suggest waterfall heights equal to or greater than 60 m. This height is possibly one of the factors that induced the exceptional depths of the Fosses Dangeard (see Chapter 6 of this dissertation). Another factor that may have contributed significantly to overdeepen these depressions is the occurrence of a sudden breach of the dam and the generation of a megaflood. We have not been able to find direct evidence proving the occurrence of that event. However, the linearity of the northern edge of the Fosses Dangeard (i.e. the final position of the waterfall before the breach) suggests that the waterfall phase ended approximately at the same time in all locations of that edge, which is consistent with a sudden breach of the ridge.

Based on the hypothetical height of the Weald–Artois ridge mentioned above and the isostatic depression the North Sea and surrounding lands may have been subjected to during the Elsterian glacial maximum, Gibbard and Cohen (2015), among others, proposed that half of the present-day surface of Belgium and almost all of The Netherlands were inundated by that lake (see Figure 2.5a; Chapter 2). However, no evidence has been found in Belgium or in the Netherlands supporting that hypothesis. The extent of that lake is therefore still debated. However, until evidence showing otherwise is found, we will use that lake extent for our palaeogeographic reconstructions (see Figure 8.1a), as that is the only estimation currently published.

The incision of the Fosses Dangeard indicates that, at the time of their formation, a significant amount of water spilled over the ridge damming the southern North Sea lake at the Dover Strait. That water most likely formed a stream, running southwestward from the waterfall and discharging into the Channel River. That means that a certain amount of fresh water coming from the southern North Sea drainage systems should have reached the Atlantic Ocean through the English Channel from the moment the lake overtopped the Weald–Artois ridge. The eventual breach of the "Dover-Strait dam" would have resulted in a dramatic increase of the volume of water flowing through the English Channel. The possible megaflood generated during the breach and the major river possibly established afterwards most likely caused intense erosion in the English Channel, incising some of the palaeovalleys of the Channel palaeovalley system. After the breach, the lake outflow would thus have become the main palaeo-river running through the English Channel, into which all other British and French rivers discharged. That geographic setting would have remained until the merged ice sheets blocking the northern fluvial route of southern North Sea palaeo-drainage systems broke up.

Based on the spatial distribution of exposed and buried palaeovalleys mapped, among others, by Smith (1980), James et al. (2002), Mellett et al. (2013) and Collier et al. (2015) in the English Channel, we have tentatively reconstructed the English Channel palaeo-drainage system before and after the Weald–Artois breach (Figure 8.1). In these reconstructions, we assume that the Northern Palaeovalley of the Channel River was not active at that time. We base this assumption on models presented by Lericolais et al. (2003) and Mellett et al. (2013), suggesting that it was not until Saalian-Weichselian times that the Channel River migrated that far north.

A lake inundating the entire southern North Sea is incompatible with the continuation of the Thames–Medway palaeo-river system through the Outer Thames Estuary (OTE). If our interpretation is correct and palaeovalley Pvs6 represents the diverted Thames–Medway palaeo-fluvial system that traversed the OTE during the Elsterian glaciation, the incision of this valley should have occurred either before the ice sheets merged across the North Sea, or after they split up (see Figure 8.2), or due to a dramatic drop of the lake level induced by the breach of the Weald-Artois ridge as theorized in Figure 8.1b.

The geomorphological study of the OTE has also revealed a series of elongated palaeo-depressions (Ed3–IGD) similar to those carved by glacial erosion, suggesting that the Elsterian ice sheets may have reached this area possibly during the Elsterian glaciation (see Figure 8.1 and Figure 8.2). This prolongation of the ice sheet along the OTE may actually be the cause of the Thames–Medway diversion. More importantly, these possible glacial erosional features cast doubt on the maximum southern expansion of the British–Irish ice sheets across the southern North Sea during the Elsterian Glaciation proposed by previous studies (e.g. Lee et al., 2012).

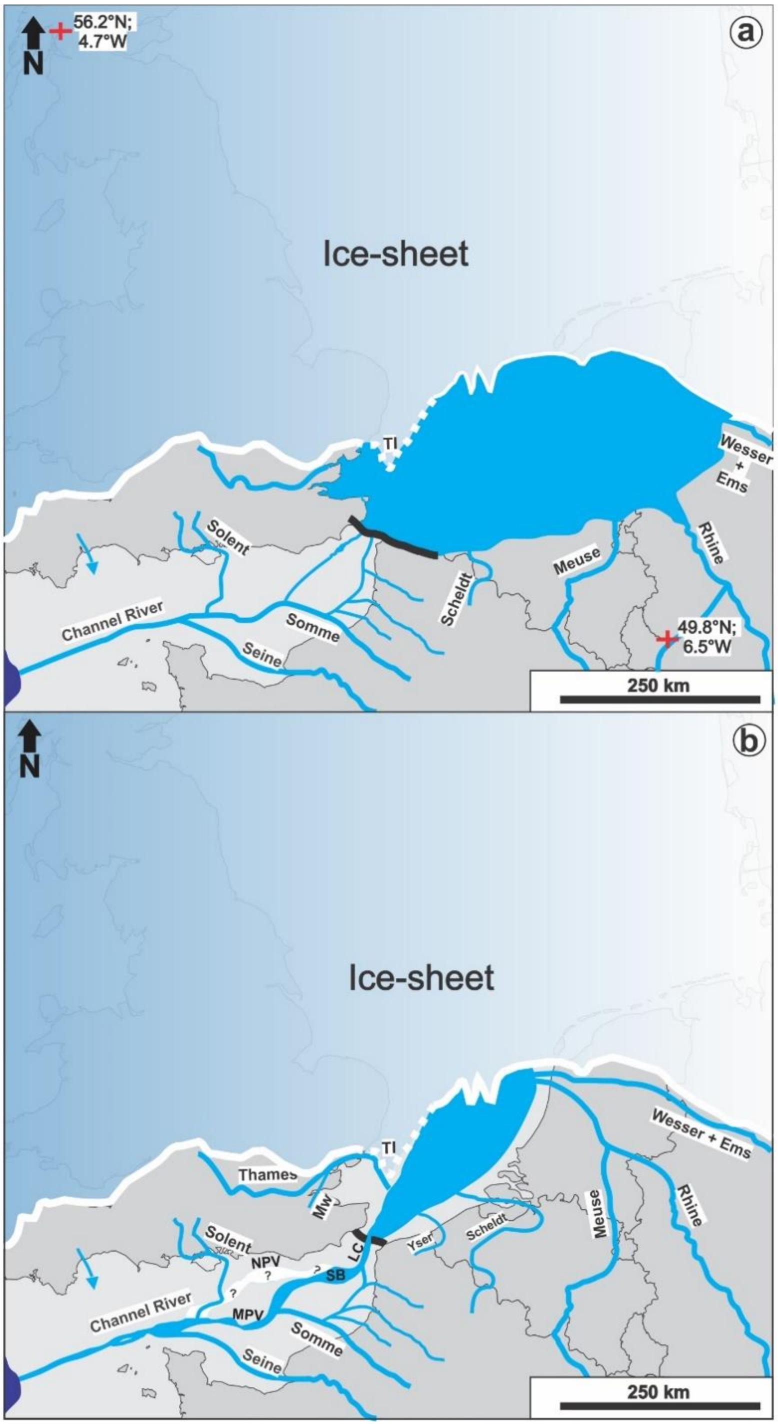

Figure 8.1. Sketch showing the hypothetical hydrographic configuration before (a) and after (b) the breach of the Weald–Artois ridge. Lake extent in (a) after Gibbard and Cohen (2015). Lake extent and palaeo-river watercourses in (b) hypothesized by the author of this study. The various palaeovalleys of the Channel palaeovalley system are indicated; these are: LC: Lobourg Channel; SB: South Basserelle palaeovalley; NPV: Northern palaeovalleys; MPV: Median palaeovalley. Mw: Medway palaeo-river. TI: hypothetical tongue of ice extending southward across the Outer Thames Estuary (OTE).

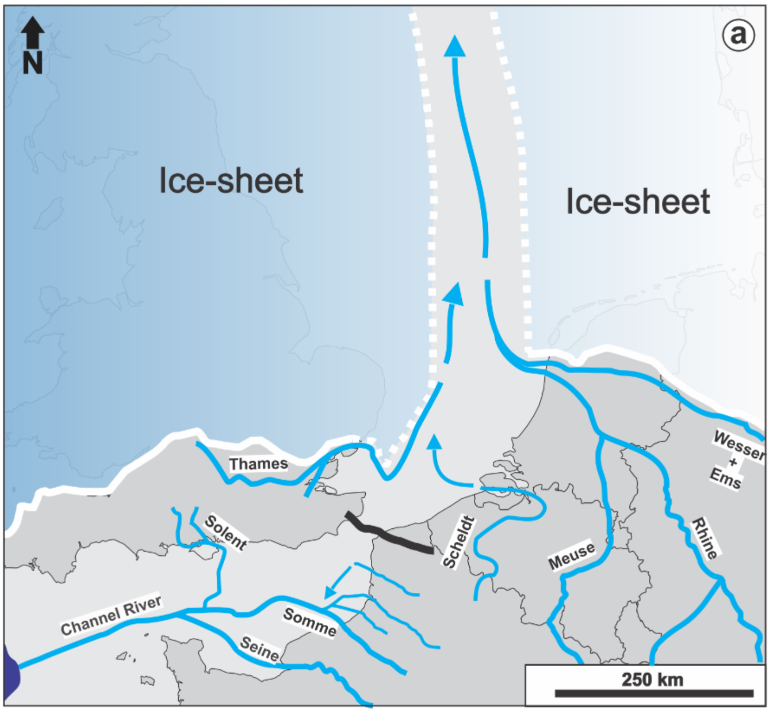

Figure 8.2. Sketch showing an alternative hypothetical scenario proposed for the southward diversion of the Elsterian Thames–Medway palaeo-river system across the OTE; i.e. before and/or after the ice-sheets merged across the central–northern North Sea.

8.2 Saalian glacial maximum (175–140 ka BP)

The morphologies and infills of the Fosses Dangeard revealed in the present study indicate the occurrence of only one episode of waterfalls in the Dover Strait. This implies that a lake was dammed at that location only once. Hence, if the Fosses Dangeard were formed during the Elsterian glaciation, other lakes inundating the southern North Sea during subsequent glacial periods did not extend to the Dover Strait. This is consistent with palaeogeographic reconstructions, suggesting that the possible lake that was formed in the southern North Sea during the Drenthe Stage (175–155 Ka BP) of the Saalian glacial maximum did not inundate the entire unglaciated area (e.g. Gibbard, 1995; Gibbard and Cohen, 2015). Rather, that lake was confined in between the southern margin of the merged ice sheets in the north, and a land-bridge that extended across the southern North Sea a few tens of kilometers to the north of the Dover Strait, in the south (Figure 8.3a). The restriction of that lake to the northern part of the unglaciated area of the southern North Sea is also consistent with studies on the Yser River (e.g. Bogemans et al., 2016). These studies suggest that no lake has extended over the Belgian western coastal plain since MIS 9 (i.e. 337–300 ka BP).

According to Gibbard and Cohen (2015), the ice-marginal lake evolved into a continuous major river system at the Warthe Stage; i.e. 150–140 ka BP. These authors hypothesized that the land-bridge damming the lake in the south may have been breached suddenly, inducing high-magnitude flood flows in the southern North Sea – English Channel (e.g. Gupta et al., 2007). We have not found direct evidence of the formation of that lake in the southern North Sea or of its sudden draining. However, the generation of highly erosive fluvial or high-magnitude flood flows in the eastern English Channel between the Elsterian and Eemian glaciation is consistent with our observations in the Dover Strait. Indeed, two internal erosional surfaces (Eb1 and Eb4) interrupting the infills of the Fosses Dangeard suggest the occurrence of intense erosional events produced by flooding and/or intense fluvial erosion. These events appear to have taken place before a marine transgression older than the Holocene, which suggest that they may have happened before the Eemian marine transgression. Intense fluvial erosion or flooding occurring prior to the Last Glacial Maximum is also attested by valley incisions LC1 and LC2. Our study therefore supports the occurrence of several extreme erosional episodes during the Saalian glaciation. These may be related to megafloods produced by a sudden emptying of a southern North Sea lake and/or other lakes discharging into the drainage system that traversed the southern North Sea at that period (see Meinsen et al., 2011)

Evidence suggesting the presence of a palaeo-river system traversing the southern North Sea remains also scarce. The geomorphological investigation of the seafloor show the existence of fluvial erosional features possibly formed during the Saalian period, such as palaeovalleys ThP1, Pvs7 and scarp S4. However, none of the palaeovalleys and apparent valleys' palaeo-slopes identified in the southern North Sea are incompatible with the presence of an ice-marginal lake as depicted in Figure 8.3b (compare with Figure 8.4).

Figure 8.3 and Figure 8.4 show tentative reconstructions of the southern North Sea hydrographic networks during the Drenthe and Warthe Stages of the Saalian glacial maximum. The distribution of channels along the Channel palaeo-river is derived from maps shown in Mellett et al. (2013). The formation of the Northern Palaeovalley during this period is still uncertain, as dating of fluvial sediments from its palaeo-watercourse yielded a Last Glacial Maximum age (Mellett et al., 2012; Mellett et al., 2013). However, the geomorphological investigation of Gupta et al. (2007) indicates the occurrence of at least two intense erosional events in that palaeovalley. Hence, until more information is available from the Northern Palaeovalley, we have tentatively included it in the palaeogeographic reconstructions proposed in this study for the Saalian glacial maximum.

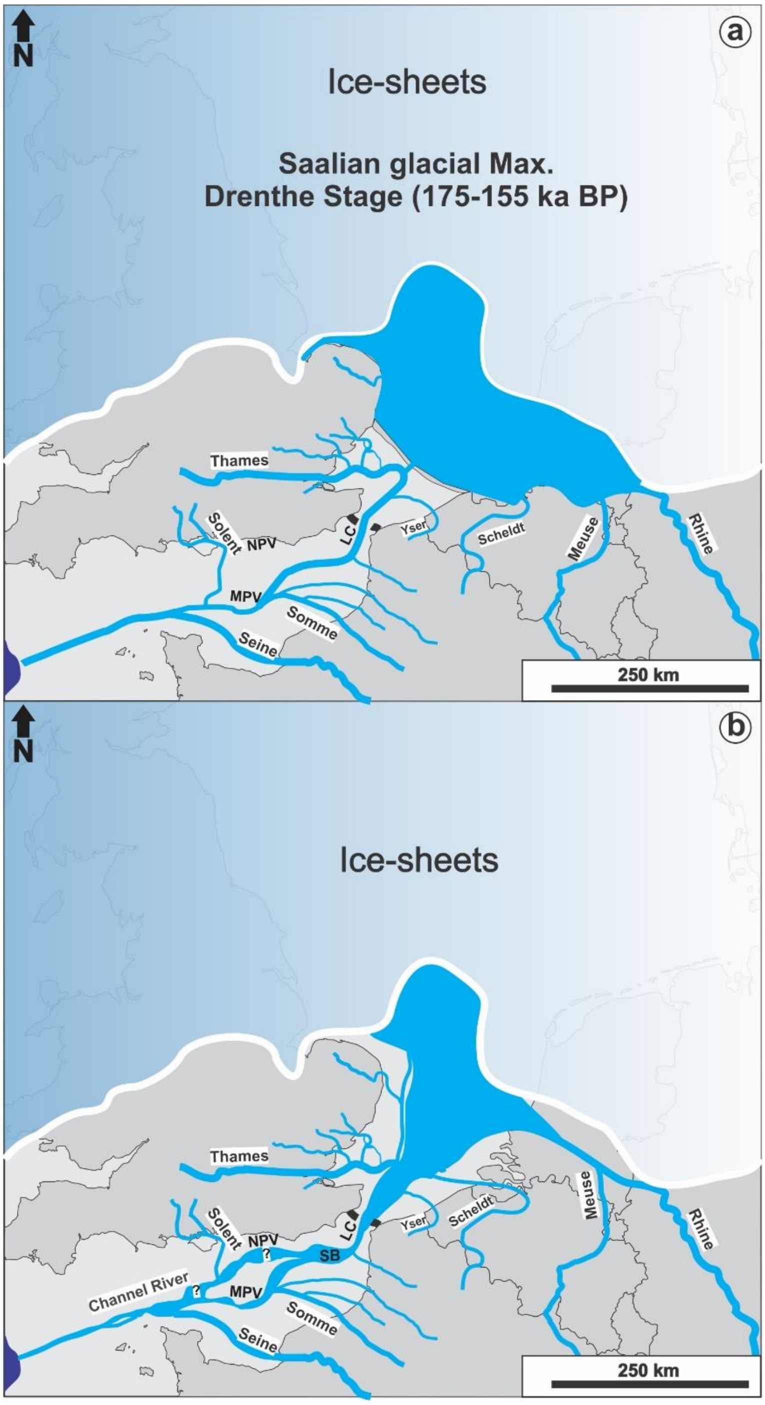

Figure 8.3. a) Sketch showing the hydrographic network during the Drenthe Stage of the Saalian glacial maximum before the draining of the southern North Sea lake. Modified from Gibbard (1988), Gibbard (1995) and Gibbard and Cohen (2015). b) Hypothetical hydrographic network after the breach of the land-bridge damming the ice-marginal lake based on the geomorphology of the English Channel and southern North Sea seafloors. River watercourses are hypothesized by the author of this dissertation from combinations of the results of this study with reconstructions from Gibbard (1988), Gibbard (1995), Cohen et al. (2014) and Mellett et al. (2013). See previous figures for other labels.

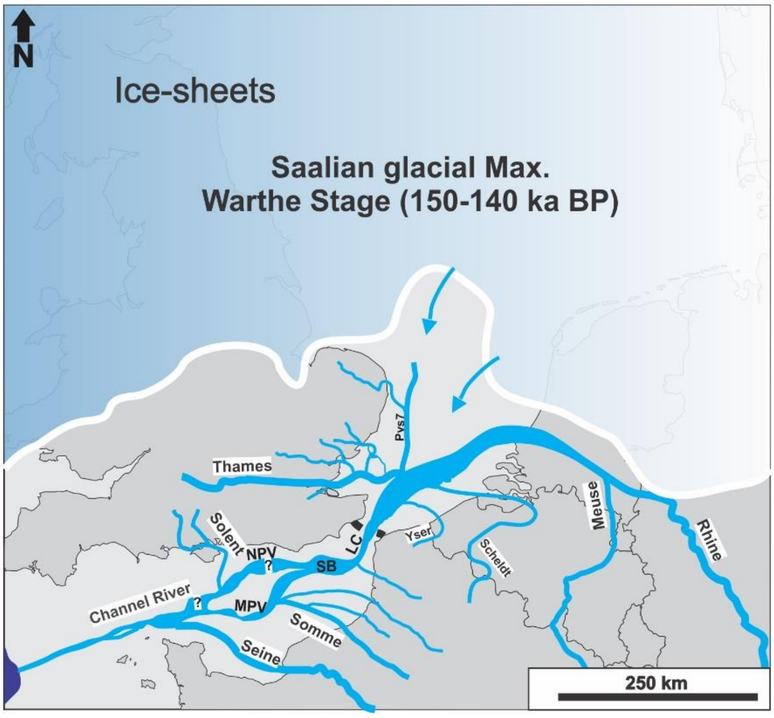

Figure 8.4. a) Sketch showing the hydrographic network during the Warthe Stage of the Saalian glacial maximum, supposing a continuous river system without development of lakes in the southern North Sea. River watercourses are hypothesized by the author of this dissertation from combinations of the results of this study with reconstructions from Gibbard (1988), Gibbard (1995), Cohen et al. (2014) and Mellett et al. (2013). See previous figures for other labels.

8.3 Late Weichselian glacial stages (30–18 ka BP)

The geomorphological analysis of the southern North Sea has revealed a series of scarps and palaeovalleys attesting to several phases of fluvial valley incision (see Chapter 7). The last phases of incision along the major palaeovalleys, such as those representing the continuations of the Thames, Rhine and Meuse Rivers, corroborate previous palaeogeographic reconstructions, proposing that those rivers were channeled through the Lobourg–Axial Channel during the Last Glacial Maximum.

The geomorphology of the seafloor also indicates that a major palaeo-river (i.e. The North Axial Channel), running southward from the German–Danish continental shelf, may have discharged into the Rhine–Meuse–Thames–Axial–Channel palaeo-river system. According to palaeogeographic reconstructions, the palaeo-river system represented by the North Axial Channel should have comprised all palaeo-rivers entering the southern North Sea from northwestern Germany and the north of The Netherlands. This river system either formed an ice-marginal lake in the Danish–German continental shelf, or converged into a major river system flowing toward the southwest (e.g. Toucanne et al., 2010; Toucanne et al., 2015; Sejrup et al., 2016; Patton et al., 2017). In any case and independently of whether the North Axial Channel was the result of a lake outflow or of converging rivers, it is clear in the bathymetry that this palaeo-river was one of the largest palaeo-drainage systems traversing the southern North Sea during the last glacial period.

The geomorphological characterization of the various palaeovalleys imprinted on the seafloor suggests the occurrence of, at least, two major phases of valley incision during the Last Glacial Maximum. The oldest one appears to correspond to the Rhine–Meuse–Thames–Axial palaeo-river system as established by models presented, among others, by Busschers et al. (2007) and Hijma et al. (2012) for the span 30–25 ka BP. The youngest valley incision, on the other hand, has not been identified prior to this study. During that phase, the erosion appears to have concentrated along the western slope of the North Axial–Axial–Lobourg Channel, resulting in the incision of the inner channel Aic–Lic in the south and NAic in the north. The erosional processes that carved these inner channels seem to have been extremely intense. For instance, the Lic exhibits erosional features similar to those found in megaflood-eroded terrain (see Chapters 5 and 6). In addition, the NAic–Aic–Lic is rather rectilinear and shows similar depth and morphology regardless whether it cuts through chalk or clays. Hence, it is unlikely that the concentration of the erosion along that channel occurred because of progressive river erosion. Alternatively, we propose that the incision of these inner channels took place during one or several megaflood event(s). The data available for this study is not enough to determine when this flooding episode(s) occurred. However, the NAic–Aic–Lic palaeovalley is younger than the palaeovalley purportedly representing the 30–25 ka Rhine–Meuse palaeo-river system. This channel should therefore have been formed during later stages of the Last Glacial Age. The source of this megaflood(s) is unknown, although it may be related to one or several of the significant meltwater discharges from the Channel palaeo-river that occurred within the intervals 22.7–21.1 ka BP, 20.5–18.4 ka BP and 18.4–16.5 ka BP (Toucanne et al., 2015). Note that once the Lic–Aic–NAic palaeovalley was formed, palaeo-rivers running into the southern North Sea and across the Dover Strait would have been channeled through it. This palaeovalley thus became the main palaeo-river of the southern North Sea palaeo-drainage system, connecting with the Channel River through the Dover Strait. It is therefore possible that subsequent progressive fluvial erosion significantly contributed to its final morphology.

We have tentatively reconstructed the southern North Sea hydrographic network during the last two phases of intense erosion identified in the southern North Sea in Figure 8.5. In these reconstructions, we hypothesize the formation of an ice-marginal lake in the north, resulting from the convergence of the German rivers Ems, Weser and Elbe. This hypothesis is based on reconstructions by Sejrup et al. (2016) and Patton et al. (2017), and the apparent lower relief of that area observed in the bathymetry compared to its surroundings. The Channel River in these reconstructions is based on the bathymetry, as well as on models presented by Mellett et al. (2013).

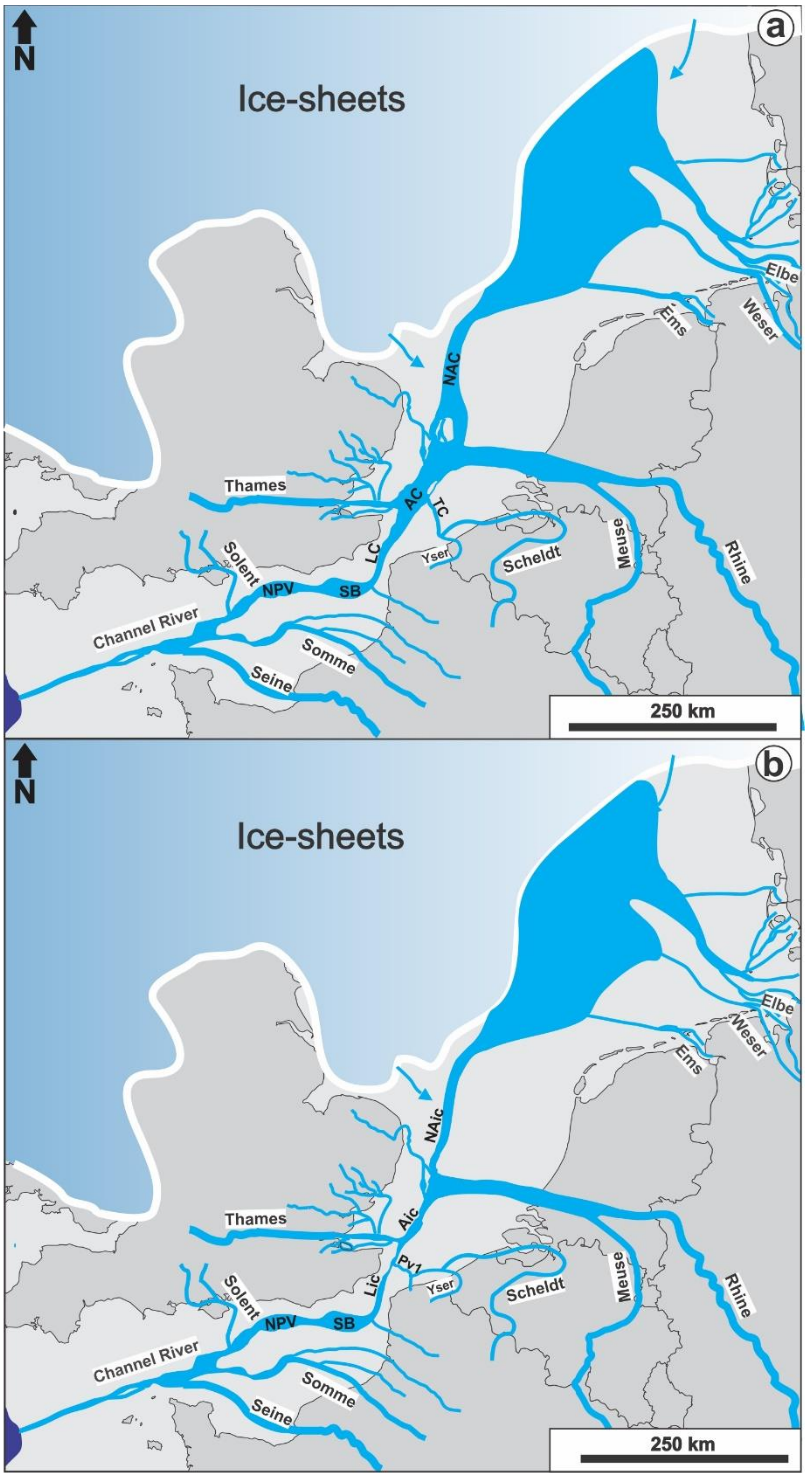

Figure 8.5. Sketch depicting the hypothetical hydrographic configuration during the two possibly Last valley incisions along the Axial–Lobourg Channel. a) Before the Lic–Aic–NAic incision; b) after the Lic–Aic–NAic incision. Note that the tributary Channel (TC) of the Axial Channel in (a) and that of the Lic (Pvs1) in (b) have been tentatively linked with prolongations of the Yser and Scheldt palaeo-rivers. This is based on combinations of the models proposed by Hijma et al. (2012) with our geomorphological study and the present-day location of those rivers.

References

Busschers, F. S., Kasse, C., Van Balen, R. T., Vandenberghe, J., Cohen, K. M., Weerts, H. J. T., Wallinga, J., Johns, C., Cleveringa, P., and Bunnik, F. P. M., 2007, Late Pleistocene evolution of the Rhine-Meuse system in the southern North Sea basin: imprints of climate change, sea-level oscillation and glacio-isostacy, Quaternary Science Reviews, v. 26, p. 3216–3248

Cohen, K.M., Gibbard, P.L., and Weerts, H.J.T., 2014, North Sea palaeogeographical reconstructions for the last 1 Ma, Netherlands Journal of Geosciences – Geologie en Mijnbouw, v. 93, p. 1–23.

Collier, J. S., Oggioni, F., Gupta, S., García-Moreno, D., Trentesaux, A., De Batist, M., 2015, Streamlined islands and the English Channel megaflood hypothesis, Global and Planetary Change, v. 135, p. 190–206.

Gibbard, P.L., 1988, The history of the great northwest European nvers during the past three million years. Philosophical Transactions of the Royal Socity B, v. 318, p. 559–602.

Gibbard, P.L., 1995, The formation of the Strait of Dover, Geological Society, London, Special Publications, v. 96, 15–26.

Gibbard, P.L., and Cohen, K.M., 2015, Quaternary evolution of the North Sea and the English Channel, Proceedings of the Open University Geological Society, v. 1, p. 63–74.

Gibbard, P.L., and Lewin, J., 2016, Filling the North Sea Basin: Cenozoic sediment sources and river styles, Geologica Belgica, v. 19, p. 3–4.

Hijma, M. P., Cohen, K. M., Roebroeks, W., Westerhoff, W. E., and Busschers, F. S., 2012, Pleistocene Rhine–Thames landscapes: geological background for hominin occupation of the southern North Sea region, Journal of Quaternary Science, v. 27, p. 17–39.

James, J.W.C., Guennoc, P., Harrison, M., Le Bot, S., Philpott, S., Vinchon, C., Bee, E., Simien, F., Janjou, D., Garlan, T., Trentesaux, A., Mahieux, G., Briet D. and, Augris C., 2002, GEOSYNTH: A synthesis of the geology and sediments of the Dover Strait and its hinterland. CR/02/078, CDROM, British Geological Survey.

Lee, J. R., Busschers, F. S., and Sejrup, H. P., 2012, Pre-Weichselian Quaternary glaciations of the British Isles, The Netherlands, Norway and adjacent marine areas south of 68 N: implications for long-term ice sheet development in northern Europe, Quaternary Science Reviews, v. 44, p. 213–228.

Lericolais, G., Auffret, J.-P., and Bourillet, J.-F., 2003, The Quaternary Channel River: seismic stratigraphy of its palaeovalleys and deeps, Journal of Quaternary Science, v. 18, p. 245–260.

Mellett, C., Hodgson, M., Plater, A., Mauz, B., Selby, I. & Lang, A., 2013, Denudation of the continental shelf between Britain and France at the glacial–interglacial timescale, Geomorphology, v. 203, p. 79–96.

Murton, D.K. and Murton, J.B., 2012. Middle and Late Pleistocene glacial lakes of lowland Britain and the southern North Sea Basin. Quaternary International, v. 260, p. 115-142.

Patton, H., Hubbard, A., Andreassen, K., Auriac, A., Whitehouse, P. L., Stroeven, A. P., Shackleton.C., Winsborrow, M., Heyman, J., and Hall, A. M., 2017, Deglaciation of the Eurasian ice sheet complex, Quaternary Science Reviews, v. 169, p. 148–172.

Sejrup, H. P., Clark, C. D., and Hjelstuen, B. O., 2016, Rapid ice sheet retreat triggered by ice stream debuttressing: Evidence from the North Sea. Geology, v. 44, p. 355–358.

Smith, A.J., 1985, A catastrophic origin for the palaeovalley system of the eastern English Channel. Marine Geology, v. 64, p. 65–75.

Toucanne, S., Zaragosi, S., Bourillet, J. F., Gibbard, P. L., Eynaud, F., Giraudeau, J., Turon, J.L., Cremer, M., Cortijo, E., Martinez, P., and Rossignol, L., 2009a, A 1.2 Ma record of glaciation and fluvial discharge from the West European Atlantic margin. Quaternary Science Reviews, v. 28, p. 2974–2981.

Toucanne, S., Zaragosi, S., Bourillet, J.F., Cremer, M., Eynaud, F., Van Vliet-Lanoë, B., Penaud, A., Fontanier, C., Turon, J.L., Cortijo, E. and Gibbard, P.L., 2009b, Timing of massive 'Fleuve Manche' discharges over the last 350 kyr: insights into the European ice-sheet oscillations and the European drainage network from MIS 10 to 2, Quaternary Science Reviews, v. 28, p. 1238–1256.

Toucanne, S., Zaragosi, S., Bourillet, J. F., Marieu, V., Cremer, M., Kageyama, M., Van Vliet-Lanoë, B., Eynaud, F., Turon, J-L, and Gibbard, P. L., 2010, The first estimation of Fleuve Manche palaeoriver discharge during the last deglaciation: evidence for Fennoscandian ice sheet meltwater flow in the English Channel ca 20–18ka ago, Earth and Planetary Science Letters, v. 290, v. 459–473.

Toucanne, S., Soulet, G., Freslon, N., Jacinto, R. S., Dennielou, B., Zaragosi, S., Eynaud, F., Bourillet, J-F., and Bayon, G., 2015, Millennial-scale fluctuations of the European Ice Sheet at the end of the last glacial, and their potential impact on global climate. Quaternary Science Reviews, v. 123, p. 113–133.

Chapter 9

Conclusions

Chapter 9 – Conclusions

9.1 Conclusions

The geomorphological analysis of the Dover Strait and the southern North Sea has revealed the presence of a series of palaeovalleys and palaeo-depressions, demonstrating the occurrence of several subaerial erosional episodes prior to the Holocene Epoch. The various palaeovalleys and palaeo-depressions imprinted on the southern North Sea and Dover Strait resulted from multiple phases of subaerial and submarine erosion, which represents a major challenge for reconstructing their evolution over time. However, even though we have not been able to unravel the whole evolution of these features from their formation to their present-day morphology, this study provides important clues on how they originally formed.

The most significant conclusion derived from the geomorphologic characterization of the major erosional features incised in the Dover Strait is the corroboration that the Fosses Dangeard were most likely formed by plunge-pool erosion at the base of waterfalls. This study thus confirms the presence of a lake occupying the southern North Sea, at least once, during the Pleistocene Epoch. More importantly, the morphology of the Fosses Dangeard suggests that the waterfall and subsequent (possibly sudden) breach of the ridge that dammed the lake led to the initial opening of the Dover Strait. The absolute timing of these events remains poorly constrained, due to the lack of contemporaneous sediments. However, indirect evidence published in previous studies suggests that the opening of the Dover Strait (and hence the formation of the Fosses Dangeard) may have occurred around 450 ka ago; i.e. during the Elsterian glacial maximum.

Incidentally, by studying the geology and tectonic structure of the Dover Strait in order to assess the processes that formed the Fosses Dangeard, we have reached the conclusion that neither the Fosses Dangeard nor the Dover Strait were formed by tectonic forcing. Indeed, our study demonstrates that tectonic activity has been very low during the Quaternary Period. The characterization of the fault activity in this area has also led to the construction of new and more detailed geological and structural maps of the submarine North-Artois Shear Zone in the Dover Strait.

Another contribution of the present study is the geomorphological characterization of palaeovalleys and palaeo-depressions incised in the Outer Thames Estuary. This has led to the identification of several palaeovalleys never described before, as well as to the corroboration of the southward diversion of the Thames–Medway palaeo-river system during the Elsterian glacial maximum. However, the latter has to be validated by dating, as it is based on previous palaeogeographic reconstructions. The geomorphology of a series of isolated deeps suggest that the diversion of that palaeo-river system may have been induced by an extension of the British–Irish ice sheets, which appears to have extended up to the Outer Thames Estuary at some point during the Elsterian glaciation. This is in contradiction with previous reconstructions of the maximum extents of the ice sheets during that glacial period, suggesting that the southward extent of the ice sheet during the Elsterian glaciation should be revised.

The geophysical data available for this study show the existence of two major phases of valley incision along the Thames palaeovalley system. Based on their infills, location and interrelationship, we interpret these palaeovalleys as the eastern continuations of the Thames palaeo-river during the Saalian and Weichselian glacial periods, respectively.

We have also found evidence in the seafloor and in the infill of the Fosses Dangeard suggesting the occurrence of, at least, two megafloods and/or high-energy fluvial erosions that may have occurred during the Saalian glacial maximum. This supports previous hypotheses postulating a catastrophic or rapid erosion of the land-bridge that purportedly dammed a proglacial lake in the northern part of the unglaciated area of the southern North Sea 175–155 ka ago.

The Weichselian glacial maximum appears to be the period that is best expressed in the geomorphology of the southern North Sea. We have identified a prominent NE–SW-oriented palaeovalley in the western part of the southern North Sea, exhibiting two major phases of valley incision that possibly occurred during that glacial stage. This palaeovalley corresponds to the Axial Channel, into which the Rhine–Meuse and Thames palaeo-river converged. Our investigation shows that this palaeovalley system is joined by a palaeovalley, in this study referred to as the "North Axial Channel", which extends northward up to the possible location of a major palaeo-river that purportedly ran southward from the German–Danish continental shelf 30–18 ka ago. This corroborates previous hypotheses suggesting that major German palaeo-rivers converged into an ice-marginal lake or a major palaeo-river on the German–Danish continental shelf of the southern North Sea owing to the ice blocking their northern routes. The geomorphology of the southern North Sea suggests that the outflow of that lake or the southwestward continuation of that river system flowed toward the Dover Strait.

Prominent inner channels carved within the Axial Channel and North Axial Channel correlate with the Lobourg inner channel (Lic) in the Dover Strait, which exhibits erosional features typical of megaflood-eroded terrains. This suggests the occurrence of one or several megaflood events during the Last Glacial Period. The absolute timing of these events is unknown. However, based on palaeogeographic reconstructions presented by Toucanne et al. (2015), Hijma et al. (2012) and Sejrup et al. (2016), and the fact that these inner channels represent the last major erosional episode imprinted on the bathymetry, we propose that the megaflood(s) that carved them happened during the Last Glacial Maximum and/or subsequent massive meltwater discharges (i.e. 30–16 ka BP).

In conclusion, the present study provides key information on the major palaeogeographic modifications the southern North Sea and English Channel went through during the last half a million years. It also provides geomorphological evidence corroborating and/or complementing some of the hypotheses previously postulated on the opening of the Dover Strait and the configuration of the southern North Sea hydrographic network during Middle–Late Pleistocene glacial maxima. Last but not least, the geomorphological investigation presented in this dissertation indicates that previous models of the ice-sheet southern extents and on the distribution of rivers and lakes in the southern North Sea should be revised.

9.2 Recommendations for future works

The palaeogeographic and hydrographic reconstructions derived from combinations of the results presented in this study with previous palaeogeographic investigations remain incomplete owing to the lack of absolute dating and a more extensive geophysical/geological database. Here below, we list a series of propositions for future works that may solve some of the open questions identified in this dissertation.

1. A geological campaign to collect sediment cores from the Fosses Dangeard to perform sedimentary characterization and age dating of their infills. Especially important are the ages of the various units infilling Fosses A and B. Indeed, these may not only give a minimum age for the formation of the Fosses Dangeard, and so of the initial opening of the Dover Strait, but they may provide valuable information on the

evolution of this area in the span between the Elsterian glaciation and the Eemian interglacial. In addition, the age of their uppermost infill will provide a maximum age for the megaflood that incised the Lobourg inner Channel.

2. Acquisition of geophysical and sedimentary data along the North Axial Channel, and its possible intersection with the lake (or major river) in the north, and with the Axial–Rhine–Meuse system in the south. This would prove or disprove the presence of a lake in the north and whether or not it produced the megaflood that carved the NAic–Aic–Lic, as well as confirm the age of the Lic and provide information on the last phases of evolution of the southern North Sea during the Last Glacial Age.
3. Acquisition of geophysical and sedimentary data from the area between the Belgian Continental Shelf and the Dover Strait. This may provide information on the ages of the escarpments carved in the Dover Strait and their relationship with the ones observed in the Belgian Continental Shelf, which is key to understand the evolution of the Rhine–Meuse palaeo-river system since the Dover Strait opened.
4. Acquisition of geophysical and sedimentary data from palaeovalleys Pv3 and Pvs6, and from palaeo-depressions Ed3–IGD in order to unravel the Elsterian history of the Thames palaeo-river and to corroborate the possible southward expansion of the British–Irish ice sheet along the Outer Thames Estuary during that period.
5. Acquisition of geophysical and sedimentary data from palaeovalleys ThP1 and ThP2 and their westward and eastward continuations. This would shed light on the relationship of the Thames palaeo-river with the Axial Channel during the last two major Pleistocene glaciations, as well as provide information to better understand the Quaternary evolution of these palaeo-rivers.

References

Gibbard, P.L., and Cohen, K.M., 2015, Quaternary evolution of the North Sea and the English Channel, Proceedings of the Open University Geological Society, v. 1, p. 63–74.

Hijma, M. P., Cohen, K. M., Roebroeks, W., Westerhoff, W. E., and Busschers, F. S., 2012, Pleistocene Rhine–Thames landscapes: geological background for hominin occupation of the southern North Sea region, Journal of Quaternary Science, v. 27, p. 17–39.

Sejrup, H. P., Clark, C. D., and Hjelstuen, B. O., 2016, Rapid ice sheet retreat triggered by ice stream debuttressing: Evidence from the North Sea. Geology, v. 44, p. 355–358.

Toucanne, S., Soulet, G., Freslon, N., Jacinto, R. S., Dennielou, B., Zaragosi, S., Eynaud, F., Bourillet, J-F., and Bayon, G., 2015, Millennial-scale fluctuations of the European Ice Sheet at the end of the last glacial, and their potential impact on global climate. Quaternary Science Reviews, v. 123, p. 113–133.

Annexes

Annex A – sediment cores from the central Dover Strait

The table below shows the positions of the boreholes gathered from the BRGM (see Chapter 4) in latitude/longitude (WGS84), and the depth of the top surface of each sedimentary formation measure in meters from the seafloor.

Id	Longitude	Latitude	White Chalk	Grey Chalk	Blue Chalk	Tourtia	Gault Clay	Lower Greensand	water depth
1	1.33	51.04		21	47.3	70.82	72.13	117.18	30.5
2	1.34	51.04					59.4		30.5
3	1.49	51.03	1.6	17	52	75.59	82		55.6
4	1.51	51.02	6.6	26	52	87			51.2
5	1.51	51.01			28.8	44.27	50.09		62.4
6	1.53	51.01			27.5	47.5	52.98		54.6
7	1.55	51.00			40.03	62.64	67.18		42
8	1.56	51.00				16.94	21.56		47.4
9	1.56	51.00		8.95	46.7	67.55	71		47.2
10	1.57	51.00		5	42.47	62.54	65.2		46
11	1.58	50.99			35	61.68	63.75		48
12	1.58	50.99				6.17	8.9		47.5
13	1.59	50.99		16.4	48.95	70.79	73.3		42.8
14	1.59	50.99	3	14.73	55	81.4	84.07		45.3
15	1.60	50.99			5.79	27.1	30.1	51	46.1
16	1.60	50.99		30.63	70	95.91	98.55		43.8
17	1.61	50.98				26.42	27.61		39.4
18	1.61	50.99	13.4	29	60	94.8	97.25		40
19	1.62	50.98			39	65.2	66.73		50
20	1.63	50.98		37.6	60.8	84.1	88.9	108	47
21	1.63	50.98			22	47.7	50.17		39.9
22	1.64	50.98		23.9	70	89.3	90.7		41.8
23	1.64	50.97	6.1	21.55	59.7	82.3	84.42		35.7
24	1.65	50.97			22	50.3	51.71		36.3
25	1.65	50.97		20.5	55	81.05	82.75		35.4
26	1.66	50.97				83.4	88.1	110	34.7
27	1.66	50.97			23.6	52.6	53.9		32.9
28	1.66	50.97	35		78.57	111.2	113		33
29	1.67	50.97		26	69.3		71.54		31.6
30	1.68	50.97		16.4	44	62.8	65.6	81.6	29.9
31	1.70	50.96		31		69.64	71.21		31.3
32	1.71	50.96	3.5	19.7		76.6	78.5		24.2
33	1.72	50.96	101.5			124	126.4		23.1
34	1.71	50.96		17	46	62.6	64.9		22.9

35	1.72	50.96		22.5	34	56.3	59		19.4
36	1.73	50.96	34.87	52.45	77.9		101.63		21.6
37	1.74	50.95	29	46.93	68.7	86.2	88.73		15.5
38	1.46	51.03			30	60	67.8		55.5
39	1.45	51.05		10	48.5	72.3	77.1		52.8
40	1.44	51.07	10.29	26.72	60	85.17	87.53		34.4
41	1.43	51.07	12.04	28.35	60	68.88			34.7
42	1.42	51.07	27.89	42.67	68.7	103.38	107.29	146.57	8
43	1.41	51.08	1.22	21.6	50	77.88	79.81		32
44	1.37	51.11		4	39.5	58.27	59.53		35
45	1.35	51.11		9	40	66.3	67.5	113	28.6